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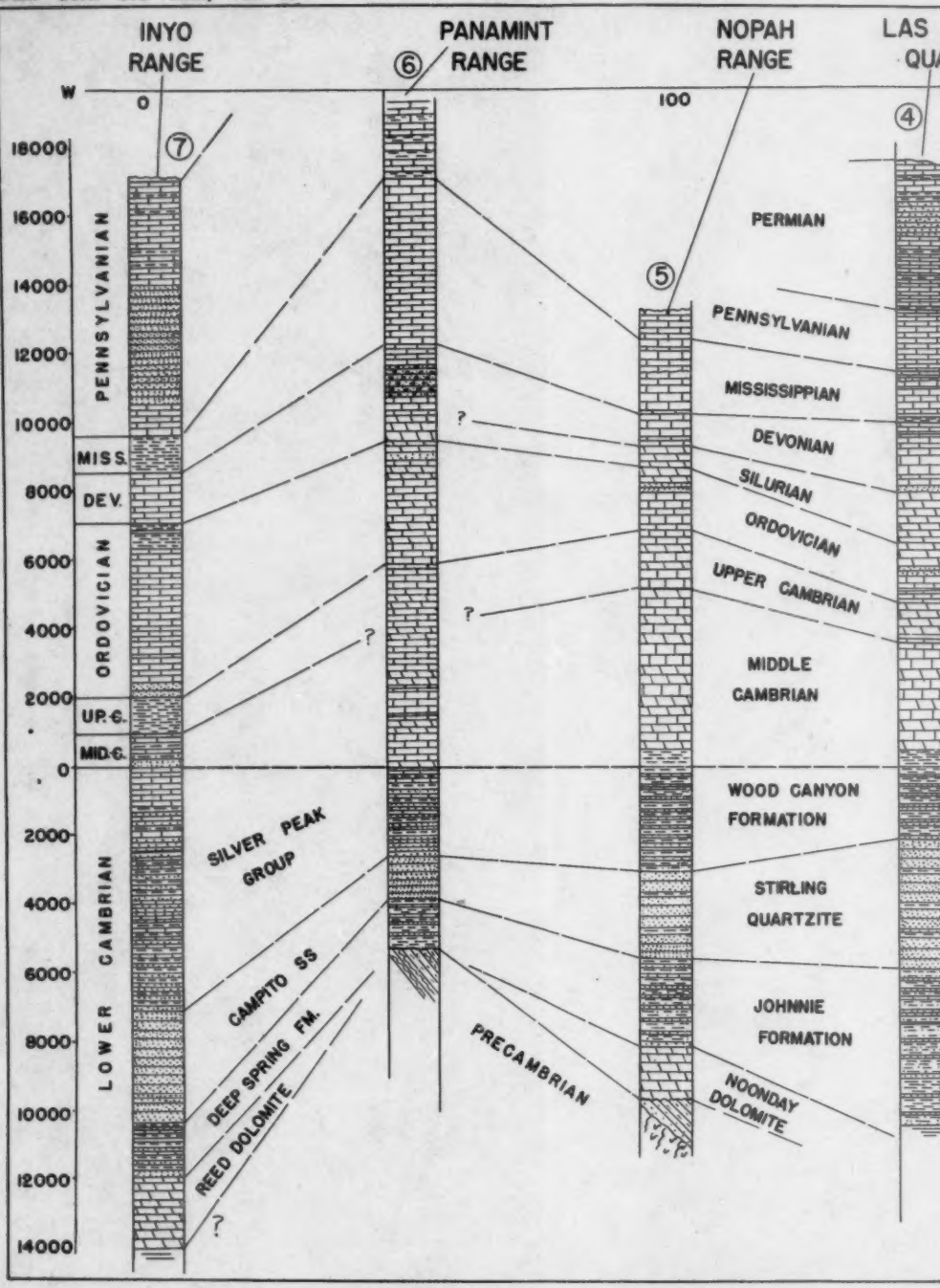
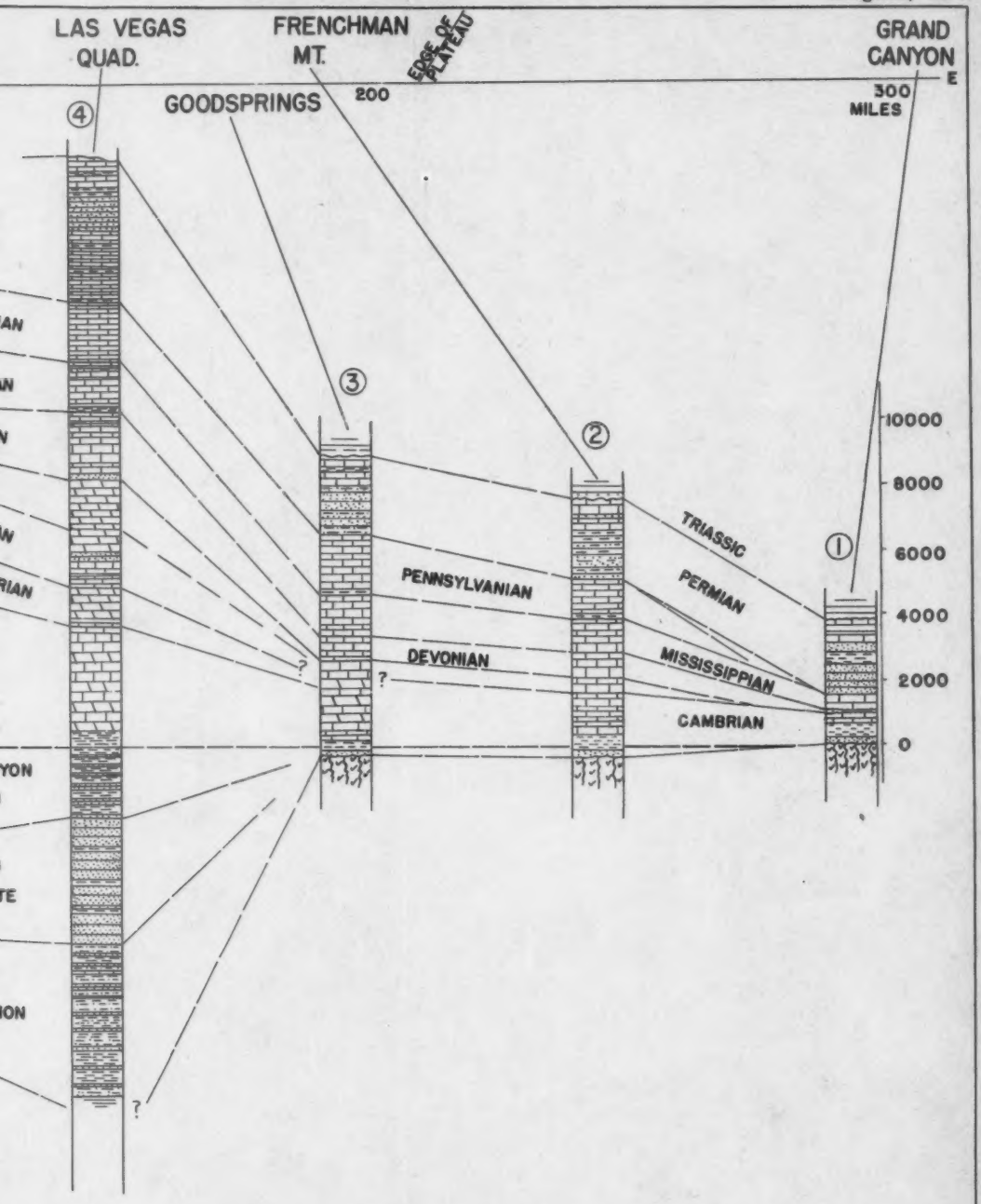


PALEOZOIC SECTIONS BETWEEN GRA

Uncertain correlations are indicated by question marks. Sources of information: Grand Canyon, L. F. Noble, U. S. Geol. Survey, Bull. 549, and Prof. Paper 131-B; Frenchman Mountain, C. R. Longwell, unpublished data; Goodsprings, D. F. Hewett, U. S. Geol. Survey, Prof. Paper 126; Las Vegas, C. R. Longwell, largely unpublished data, and T. B. Nolan, Am. Jour.



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Higher part of Panamint section omitted, to conserve space.

Locations of sections shown by corresponding numbers on Figure 1.

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TECTONIC THEORY VIEWED FROM THE BASIN RANGES

By Chester R. Longwell

(Address of Retiring President of The Geological Society of America)

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Introduction

One of our geologists of an older generation, widely known for his facility in deductive reasoning, was guided during his declining years to localities in the Great Basin which, without his having seen them, had played a part in his thinking and his theoretic writing. His delight was unbounded on finding that some of the features well displayed in the field were in striking harmony with deductions he had published many years earlier. There seems little likelihood, however, that any contemporary author of an ambitious hypothesis to explain the step by step development of a great mountain belt, with identification of the responsible forces and analysis of their application, will live to see his scheme crowned as demonstrated truth. In driving toward this goal our Lilliputian efforts build slowly the groundwork of fact on which an inductive edifice of theory can be erected, or to serve as a check on deductive guesses.

After 30 years in which considerable field study and thought have been devoted to some of the Basin Ranges in southern Nevada, I should be happier merely describing the features and discussing the geologic relations found in that limited area than in attempting a broad view of a great tectonic province. However, the real significance of local geology sometimes appears only when viewed within a larger framework. Conversely, theories relating to major tectonic units find some of their most crucial tests in the information furnished by units of smaller scope.

The Basin and Range province has been defined basically on topographic form (Fig. 1). It is "a region distinguished by isolated, roughly parallel mountain ranges, separated by desert basins" (Fenneman, 1931, p. 326). However, as Gilbert recognized in the course of his pioneer

work (1875), structure of bedrock has played a prominent part, both directly and indirectly, in determining the characteristic surface forms. The map, Figure 1, shows a large part of the province in the United States, in relation to the Colorado Plateaus and the Rocky Mountains on the east, and to ranges of the Pacific province on the west. Beyond the limits of this map the large fraction of the province known as the Great Basin reaches northward into southern Oregon and Idaho; and the province as a whole extends southward and southeastward far into Mexico. In this vast region of complex geology, alluvium in the wide basins, and lavas and tuffs covering large areas, present a major obstacle to satisfactory structural study. Even so, thorough field examination and mapping of the bedrock that is well exposed in the region is barely begun. Literally, therefore, for any broad view of the geology we are dependent on scattered samples. Many of these samples are magnificent displays, and supply eloquent testimony; but in attempting to integrate the evidence we must keep in mind that detailed information is available for a very small fraction of the total area, and that large gaps between localities furnishing critical data make positive conclusions hazardous.

The tectonic record of the region is written in sedimentary deposits, in deformation of earlier date than Basin Range development, as well as the Basin Range structure itself, and in igneous rocks, both intrusive and extrusive.

Record of Sedimentation

Seven sections of Paleozoic formations to be considered extend generally along an east-west line from the Grand Canyon in Arizona to the Inyo Range near the Sierra Nevada, a distance of more than 300 miles (Fig. 1). Several striking facts appear in the sections (Pl. 1). Clastic deposits at the base, uniformly thin through the eastern half of the line, thicken remarkably westward, to 10,000 feet or more northwest of Las Vegas, and to the order of 15,000 feet in the Inyo (Knopf, 1918) and Silver Peak (Turner, 1909) ranges. Above these clastic deposits, the nearly continuous carbonate section is fully as remarkable, totalling more than 15,000 feet in the Las Vegas area, where it represents the time span from Mid-Cambrian to Permian.

The full stratigraphic range of the older clastic formations is not known, and correlations among some of them are uncertain. Fossils in the thin sections on the Plateaus indicate Middle Cambrian age; whereas beds of closely similar nature and thickness west of the Grand Canyon yield Lower Cambrian forms, including *Olenellus*. Thus an eastward onlap of considerable time significance is indicated. In thick sections farther west, Lower Cambrian fossils have thus far been found no lower than the Wood Canyon formation of the Death Valley region, where the Stirling quartzite and the Johnnie formation carry the conformable sequence several thousands of feet lower (Nolan, 1929; 1943; Hazzard, 1937). In the Inyo and Silver Peak ranges, *Olenellus* occurs through nearly 5000 feet of the Silver Peak group, and there are suggestions of a trilobite fauna also in the underlying thick Campito sandstone (Knopf, 1918). Some biostratigraphers tell us it is inadvisable to classify as Cambrian any strata below the lowest known occurrence of *Olenellus* (Wheeler, 1947; 1948). This view seems to treat lightly several matters of some weight, such as variations in physical conditions affecting distribution of organisms, uncertainties of preservation, and chances of discovery in a wide region still in the exploratory stage of field study. But regardless of problems in time-stratigraphic classification, surely it is of large importance for regional interpretation that the clastic deposits in these thick sections appear to represent a closely connected sequence of events, which taken together constitute a major episode of sedimentation. The deposits are dominantly clastic; they are closely conformable—any discordances that have been recognized seem to represent no more than local disturbance and erosion; there are numerous repetitions of units that are similar in texture, structure, and composition; the sections viewed together, in the gross, are logically interpreted as a thick accumulation that changes generally, though with irregularities, from overwhelmingly clastic on the east to less clastic and apparently more marine westward. To avoid awkwardness in the present discussion, the entire assemblage is called Lower Cambrian. If there are valid ob-

FIGURE 1.—RELIEF MAP

Showing part of the Basin and Range province (bounded by full lines) and adjoining tectonic units. Dashed line S. A. marks the San Andreas fault zone. Circled numbers, 1 to 7, show locations of stratigraphic sections similarly numbered in Plate 1. *C*, area affected by Cedar Mountain earthquake of 1932. *L. V.*, Las Vegas Valley. *B*, Battle Mountain. (Reproduced by permission of Ginn & Co. from *Landforms map of the United States* by Erwin Raisz, prepared for Atwood's *physiographic provinces of North America*)

jections to this as general practice, it then would seem desirable to speak of the sequence as the Las Vegas, or Nopah, or Inyo group of formations, to imply the essential unity of record in the section, regardless of the exact horizon at which *Olenellus* may have entered the scene.

The Panamint section, nearly midway between the Nopah and Inyo ranges, seems anomalous in its lack of some major units and in the much smaller total thickness of clastic deposits (Hopper, 1947). There is strong suggestion that the floor on which deposition began was a surface with large topographic irregularities. Absence of some basal units and thinning of others in the Panamint section indicate that relief of several thousand feet persisted through a considerable part of the depositional episode. Late Precambrian strata were strongly folded and faulted in the Death Valley region, and accordingly lie with strong angular unconformity beneath the younger section. Deformation of the late Precambrian rocks extended much farther east, to the Grand Canyon district, where tilted fault blocks underlie the Cambrian section. The low-relief surface of contact there probably was developed in the cycle of erosion responsible for detritus that buried the irregular surface in the subsiding belt farther west.

Two points deserve emphasis: westward increase in the proportion of carbonates, and profusion of marks of shallow-water deposition in the more easterly sections. Through thousands of feet of the Johnnie formation ripple marks are beautifully displayed, dendritic rill marks like those seen on modern beaches at ebb tide are common, mudcracks are in evidence at many horizons, and cross-bedding is a commonplace. The thick Stirling quartzite is replete with cross-beds, channel marks, and other indications of rather rapid deposition in shallow water.

In brief, then, the evidence at hand from this line of sections suggests a long regime of sedimentation, with confluent deltas and flood plains on the east grading westward into finer bottom deposits that interfingered with increasingly limy layers. The large total thickness, together with the great width of the sedimentary lens offshore and the persistent marks of shallow-water deposition, argue strongly for a slowly subsiding bottom, with deposition maintaining a profile of near equilibrium. It is notable that none of the evidence even remotely suggests a westward limit to the seaway at this latitude. In fact there is implication of a great wedge of deposit, such as Lawson (1938; 1942) has called a "marine embankment," built toward the open ocean. As time passed the land on the east grew lower, the shore line receded, and deposition of carbonates, with only brief interruptions, held sway through long geologic periods. Subsidence continued, maintaining the wide seaway while a 3-mile thickness of beds, chiefly limestone, was laid down, to make the total Paleozoic section of lithified deposits fully 5 miles thick over a vast area.

Comparison of the sections by means of diagrams showing proportions of common sediments is of some interest (Fig. 2). In these diagrams sandstone and shale are shown together, in comparison with limestone. The Plateaus, which by location and general Paleozoic history may be thought of as a typical platform, have in the Grand Canyon section 63 per cent of clastic materials, 37 per cent limestone. This ratio is nearly inverse to that in Pettijohn's diagram of the average "foreland suite" (1949, p. 462). Westward toward the edge of the Plateaus the section thickens gradually, by addition chiefly of carbonate beds. Fifty miles west of the Grand Wash Cliffs, at Frenchman Mountain, the proportion closely approximates that of the "typical foreland" assemblage; and, in the thicker section at Goodsprings, carbonates are considerably in excess. Even the very thick sections of Las Vegas and Nopah are reasonably close approximations to "foreland type," as judged by this test. The Inyo diagram is almost identical with that of the Grand Canyon, though the total thickness of section is more than 7 times as great. Thus, if the diagrams are taken at face value, we are dealing with a deposit that is geosynclincal in areal extent and thickness, but has composition strikingly similar to that ascribed to the average foreland.

Each of the diagrams averages the compositions of unlike fractions of a thick section. Selected divisions of the Las Vegas and Inyo sections are diagrammed in Figure 3. In each case the lower clastic unit taken alone may be said to approach the "geosynclinal type." In

the Las Vegas section all higher fractions are predominantly limestone; even the complete sequence from Mid-Cambrian up has well over 80 per cent of carbonates. At Inyo the increase of limestone in the lower units, as compared with Las Vegas, is conspicuous, and limestone dominates in the middle part of the section. However, for the systems Mississippian to Permian, there is a sharp return to dominance of clastic materials. This change in character of the diagram reflects crustal unrest in that span of time, which affected the western much more than eastern parts of the region.

Shall we then, following the diagrams, say that parts of the Las Vegas-Inyo deposit, particularly the lower clastic fractions, are geosynclinal, whereas other parts are of extreme foreland type? Pettijohn lists other criteria that must be considered in addition to rock types (1949, p. 455); these criteria relate to thickness, textures, structures, composition, and other characteristics of a deposit. In two particulars only—great thickness and the presence at some horizons of considerable feldspar—do the Las Vegas-Inyo sections qualify as "geosynclinal or orogenic"; in all other respects the listed criteria make them of "platform or epeirogenic" type. To most geologists, however, these sections are part of the Cordilleran geosyncline. This is merely one example of contradictions that are encountered increasingly in the literature on thick sedimentary deposits. What are the essential qualifications of a geosyncline?

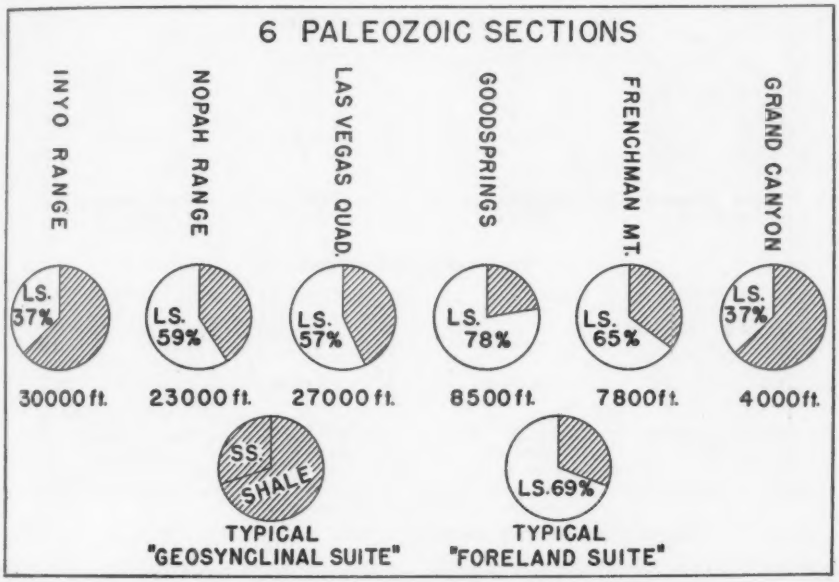


FIGURE 2.—PROPORTIONS OF COMMON SEDIMENTS IN SIX OF THE SECTIONS REPRESENTED IN PLATE 1
Diagrams of "typical suites" from Pettijohn (1949).

Actual examples that have been admitted to the genus, or family, *geosyncline* differ among themselves—in fact, each individual has some features that seem peculiar—and there is a tendency to set up classifications that grow increasingly more complicated. The difficulty of multiple classifications is aggravated by a swelling terminology that bristles with Greek appendages. Some of the discourses on this subject seem to be primarily exercises in word building. Without turning a page of one such article we find definitions for ten distinct kinds of geosynclines, with as many elaborate names that vary in length from five to eight syllables each. If we try to keep in mind also earlier—and somewhat simpler—proposed terms such as *mono-*, *poly-*, *meso-*, *oro-*, and *parageosynclines*,

and strange names for special tectonic units that lie in front of, behind, or within one or another type of geosyncline, exposition turns to confusion. The harried geologist, after some have any structural separation—one facies may simply grade laterally into the other. Are we then to apply the good term *geosyncline* to a mass of sediment whose base had no semblance

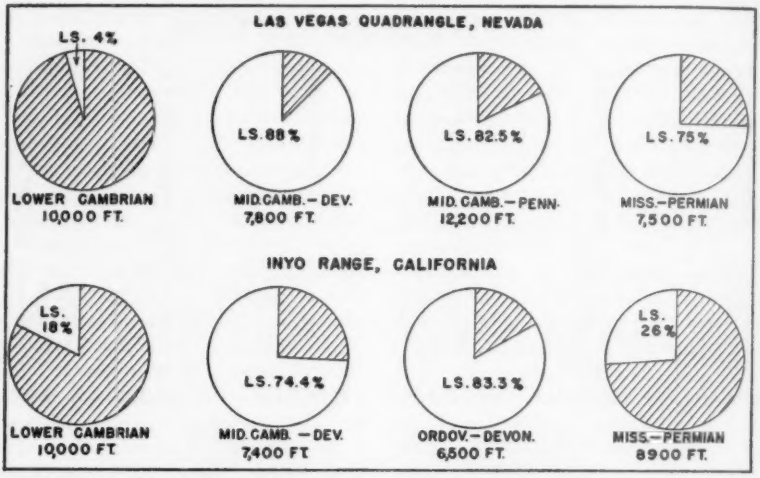


FIGURE 3.—PROPORTIONS OF COMMON SEDIMENTS IN SELECTED FRACTIONS OF THE LAS VEGAS AND INYO STRATIGRAPHIC SECTIONS

Diagonal lining represents sandstone and shale combined, as in Figure 2.

pages set with these polysyllabic neologisms, is likely to give up in despair and devote his limited time to more attractive reading.

It is unfortunate that the phalanx of cumbersome names proposed for belts of thick sedimentary deposits tends to obscure some excellent ideas. As Stille (1941), Kay (1944; 1947), and others have pointed out, there is good evidence for the concept that each of the great geosynclinal belts near a continental margin was composite, with an outer highly complex zone in which volcanic products and associated sediments accumulated; and a more or less parallel inner member, along the continental border, which received at least a part of its load of clastic sediments from the continental mass itself. The term suggested for the first is *eugeosyncline* (that is, "truly a geosyncline"); for the other, *miogeosyncline*, or "less of a geosyncline." Since neither of these terms has more than six syllables, we may consider accepting them, provided they are not to bring a dozen or more uncouth satellites in their train. But careful analysis fails to resolve some doubts. We are told that the eu- and miogeosynclines need not of synclinal form? Surely the originator of the term would not have agreed to this.

For sedimentary accumulations that have the two distinct facies in consistent relation to a continental margin, as appears to be generally true of Paleozoic deposits in the Appalachian region, may we use the homely designation *marginal* geosyncline, with *outer* and *inner* belts, or *volcanic* and *avolcanic* members? At any rate the terms adopted should be as simple and expressive as possible; not compounded with abstruse prefixes that are sure to acquire vague and varied interpretations, until the terms fade from use and become part of the wreckage with which our literature is strewn.

Many fundamental questions about Paleozoic sedimentation in the Basin and Range country cannot as yet be answered. Ideas that have been held on the early paleogeography of the Cordilleran region differ in detail but follow a general pattern. Figure 4, representing a view commonly presented regarding the Cordilleran geosyncline during Cambrian time, has the average western boundary of Cambrian seas crossing the California-Nevada State line pre-

cisely at the location of the Inyo and Silver Peak ranges, where the total thickness of lower clastics and higher Cambrian strata is fully 16,000 feet. Perhaps the intent of the map postulated shore line is convincing without clear evidence that sediments were derived from a western landmass. If the way were open as shown in this map, it may have been open much

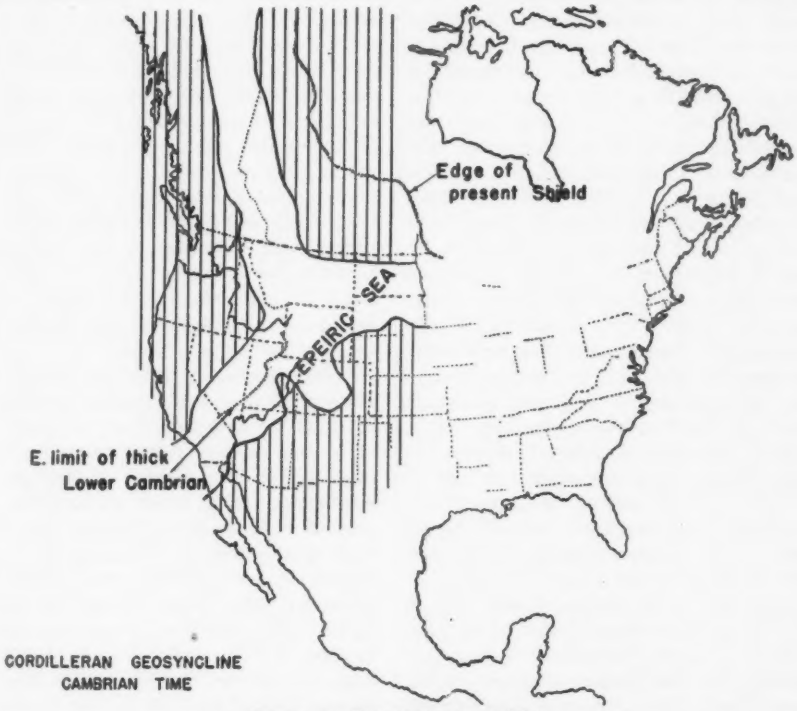


FIGURE 4.—PALEOGEOGRAPHIC MAP

Based on views commonly held regarding the Cordilleran geosyncline in Cambrian time. Supposed land areas indicated by vertical lining. Conditions in eastern North America not represented. (Taken in part from Deiss, 1941.)

makers has been to appear conservative by placing the hypothetic boundary no farther west than the tangible evidence extends; but isn't it a rather radical concept that a deposit 3 miles thick, most of it fine-grained detritus and the rest carbonates, accumulated through long geologic epochs directly at the edge of a landmass with a persistent shore line? All maps of the Cordilleran geosyncline that have current acceptance represent, without any direct stratigraphic evidence, a portal crossing a wide belt in California to provide a marine connection. These maps agree, however, in drawing a continuous western limit of the Cambrian seaway *north* of the assumed portal. But no location of a more extensively. What then became of the possible westward extensions of early Paleozoic formations? Perhaps they were destroyed or hopelessly obscured in the turmoil of orogeny, igneous activity, erosion, and sedimentation that has characterized the wide coastal belt during a long span of geologic time. Schists and other metamorphic rocks derived from sedimentary formations, Carboniferous and perhaps earlier, have been invaded by plutonic masses of the Sierra batholith (Knopf, 1929), and also in parts of the Coast Ranges (Trask, 1926; Reed, 1933). Possibly evidence eventually will reveal some units of these complexes as remnants of the missing early Paleozoic section.

Meanwhile, on what basis can the paleogeographer draw a confident boundary of the early Cordilleran sea 50 or even 100 miles west of the Inyo?

Farther north the eastern limit of thick Lower Cambrian clastic deposits is traced into southeastern Idaho (Fig. 4). To the west, in central and northern Nevada, the thick sections disappear beneath a wide cover of younger sedimentary rocks and volcanics. Surely the Cambrian formations extend much farther west than indicated in the map, Figure 4, but how much farther is a problem. The question of *Cascadia*, the supposed western source land, is involved; and I have found the search for tangible evidence on this problem fascinating but baffling. Some facts in physical stratigraphy, at Middle Cambrian and higher horizons, suggest a western source at several widely separated localities. However, each occurrence may represent only a local and temporary source—an island, or island group. The evidence is hardly sufficient to establish the concept of a large Paleozoic land that reached far east of the present Pacific coast in the United States, though recent field studies in British Columbia indicate that a landmass persisted there from Mid-Cambrian to late Paleozoic time (Armstrong, 1946; 1949; Raymond and Willard, 1931). If that land extended south of the international boundary, the evidence is effectively hidden by the wide blanket of younger volcanic rocks in the Columbia Plateau—Cascade region.

Ideas on the Paleozoic geography in the coastal belt are nebulous. In particular, we should like to know at what date volcanic activity and crustal unrest began in that belt. In northern California a thick Paleozoic section, spanning the periods Silurian to Permian, is made up chiefly of clastic deposits and contains abundant volcanic material (Diller, 1908). At least as early as Silurian time, therefore, conditions of sedimentation were very different between the Basin and Range country and the belt west of it. In the former area there was a wide seaway with chiefly carbonate deposits, whereas the coastal belt had land near by to supply abundant detritus. When did that land, and the associated volcanoes, first appear as actors in the Cordilleran cast?

One assumption may be that crustal buckling occurred west of the present coast, with attendant volcanic activity and growing island chains; and that a secondary effect was broad downwarping inland, in a belt including much of the Great Basin. Since the lack of factual evidence for this order of events seems to be total, there is room for a rival hypothesis, as follows: At the beginning of Cambrian time, or somewhat earlier, epeirogenic downwarping was initiated along the west border of the continent. This movement was contemporaneous with or followed the widespread block faulting and warping that affected the late Precambrian rocks in the Colorado Plateaus, the Death Valley region, and elsewhere. Broad upwarping inland accompanied the marginal subsidence, and a thick deposit of clastic sediments was spread seaward. Islands that projected from the irregular bottom contributed detritus locally until they disappeared under the thickening fill. The action was generally like that in the building up of the shelf deposit off the present Atlantic coast (Fig. 5). Sediments for the latter deposit have been derived chiefly from the Appalachian region, which was arched up repeatedly in Mesozoic and Cenozoic time. Ewing's seismic profiles, which he has kindly allowed me to use, indicate a present thickness of about a mile near shore, increasing at first rapidly, then more gradually and irregularly, to about 3 miles just inside the continental slope. The seismic data reveal that near the seaward limit of the thick lens the floor bends up, suggesting some degree of adjustment to the sedimentary load.

The early Paleozoic lens of sediments in the Great Basin region was much wider than the present Atlantic shelf deposit, and presumably attained greater thickness in its outer portion. Let us assume that this Paleozoic load also caused depression of the crust beneath it, and a bulge by upbending farther seaward. This bulge may have localized both orogenic failure and volcanic activity. Eventually chains of islands, like some of those off the east coast of Asia, resulted. Detritus from these landmasses, added to large volumes of volcanic materials, built up thick deposits in adjacent troughs and basins. Repeated orogeny deformed these accumulations, and also the outer part of the embankment built out from the mainland. Deformation

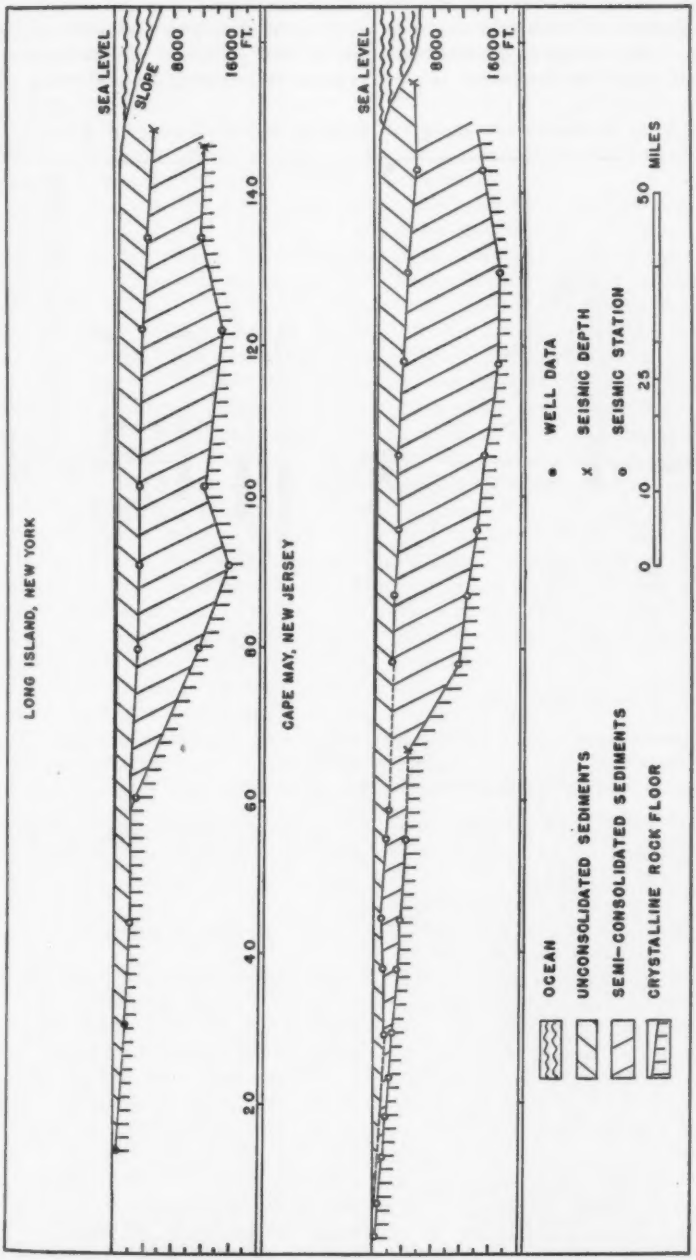


FIGURE 5.—PROFILES ACROSS THE ATLANTIC COASTAL BELT AND CONTINENTAL SHELF
Based on seismic refraction work by W. M. Ewing and associates

finally erected a partial barrier between the two contrasting theaters of deposition: an inner quiet seaway, where through long periods carbonates were the chief accumulations; and an outer chaotic belt in which sediments were chiefly clastic, in part very coarse, locally thick but in troughs with shifting axes.

Either with a sequence of events as outlined in this hypothesis, or on the assumption that the outer belt of disturbance and volcanism was first to develop, the wide inner belt of sedimentation was in the end a geosyncline as appraised by areal proportions, by cross-sectional form, and by thickness of deposits. However, the history of its growth was characterized by prolonged epeirogeny in a large part of its extent; this history is faithfully reflected in the nature of the sediments. On the other hand, some thick deposits elsewhere, such as Paleozoic formations of southeastern Alaska (Buddington and Chapin, 1929), and the late Paleozoic of the Ouachita Mountains, also qualify as geosynclinal, but in large part they reflect vigorous crustal movement near by. Therefore either *orogenic* or *epeirogenic* sediments may be also *geosynclinal*; proposed classifications of sedimentary facies that make synonymous orogenic and geosynclinal, epeirogenic and platform (or foreland), do not stand the test of field relationships and should be revised.

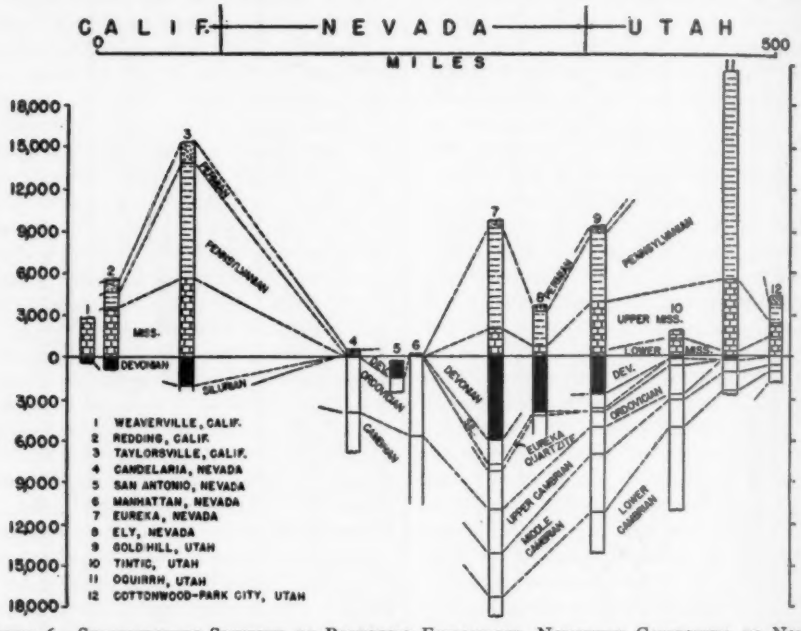


FIGURE 6.—STRATIGRAPHIC SECTIONS OF PALEOZOIC FORMATIONS, NORTHERN CALIFORNIA TO NORTH-CENTRAL UTAH

(Diagram modified from Nolan, 1928.)

RECORD OF DEFORMATION

The earliest known large-scale disturbance of the geosynclinal formations in the Great Basin area began in late Devonian time and was recurrent through the Permian. Evidence for this movement was first pointed out by Nolan (1928), after a comparison of known stratigraphic sections along an irregular line extending from northern California to northern Utah (Fig. 6). Plotting these sections with the base of the Mississippian as a horizontal line brings out graphically the location of a wide area of persistent uplift, toward which all systems from Silurian through Pennsylvanian

wedge out from both east and west. The relationships are represented more pictorially in schematic cross sections (Nolan, 1928, Fig. 3), one for the end of Devonian, the other for the end of Permian sedimentation. The uplift divided this part of the geosyncline into eastern and western troughs during Mississippian and Pennsylvanian time, and the sedimentary records on the two sides have very different aspects.

Nolan's diagrams represent the Cambrian and Ordovician systems wedging out near the Nevada-California line. This is based not on visible evidence of westward thinning but on absence of known Cambrian and Ordovician formations farther west. Eardley (1947) has modified the sections to suggest possible correlations with units of the metamorphosed series in the southern Klamath Mountains (Hinds, 1939). The proposed equivalences are frankly hypothetical. Perhaps both interpretations have merit; the formations may have thinned westward, on a continental slope, and this seaward section may form part of the metamorphosed sequence in the Klamath area.

The late Paleozoic disturbance in Nevada was at first interpreted as a broad geanticlinal uplift, involving no acute deformation. However, recent work in north-central Nevada reveals effects of severe orogeny, with development of large-scale thrusts and associated folds. Near Battle Mountain (Fig. 1), which lies on or very near the estimated crest of the uplift, thrusts of large magnitude developed before the Des Moines epoch of the Pennsylvanian period (Roberts, 1949). In the same area sharp folding affected Permian and older formations. It is highly probable that further evidence of the Paleozoic orogeny will be found in other ranges along this belt when they are carefully studied and mapped. Much farther south, in the Hawthorn and Tonopah areas, Permian sandstone rests on Ordovician with marked angular unconformity (Ferguson and Cathcart, 1924).

Late Paleozoic folding and thrusting occurred also in northern California (Diller, 1908); and since the Calaveras and associated rocks in the Sierra Nevada belt were deformed and metamorphosed before the Mesozoic folding (Knopf, 1929) presumably there was late Paleozoic orogeny of large proportions in the region. We do not know whether the deformation covered the belt, 200 miles wide, between Battle Mountain and the Sierra Nevada, or was restricted in the Great Basin to the geanticlinal area. In any case this is the earliest known extension so far eastward of orogeny generally contemporaneous with important deformation in the coastal belt.

Near the end of the Paleozoic era the crest of another north-south geanticline emerged in eastern Nevada. During Triassic and Jurassic time this uplift formed a barrier between seaways that received unlike deposits (Nolan, 1943; Schuchert, 1923). The earlier uplift to the west had disappeared, and on its site were deposited thick sections of Triassic and Jurassic strata supplied by a source to the west. Triassic rocks of that area include tremendous thicknesses of lavas and tuffs (Knopf, 1918; Muller and Ferguson, 1936). Major thrusting and folding occurred in both early and late Jurassic time, in west-central Nevada, approximately in the area of Paleozoic disturbance. The Jurassic orogeny was in part essentially contemporaneous with the folding of the Sierra Nevada-Klamath belt, which is recognized as the major deformation of the western Cordillera.

Thus field data now in hand indicate that the thrusts and folds west of central Nevada developed partly in late Paleozoic time, partly in the Jurassic period. The Paleozoic orogeny probably created highlands of at least moderate altitude, which furnished large volumes of clastic sediments in the area of central Nevada, and in similar deposits exposed in the Inyo Range. Late Jurassic deformation presumably resulted in highlands from which nonmarine sediments were spread out eastward, as represented in Reeside's isopach map for the Lower Cretaceous (Fig. 7). Groups of thrust faults in a wide strip extending from southern Nevada northeastward to the Wasatch Range are at least in part of Cretaceous age (Fig. 8). Although few of these thrusts have been closely dated, they lie in the belt of Cretaceous highlands that supplied the immense volume of detritus to the Rocky Mountain seaway, as the sedimentary evidence in central Utah so eloquently testifies (Spieker, 1946; 1949). In western Utah at least two thrusting episodes occurred before the Eocene (Nolan, 1935). In

southern Nevada one thrust developed before and another after early Upper Cretaceous time (Longwell, 1949). However, a number of the larger thrusts are not known to involve formations younger than Middle Jurassic; therefore caution is required in assigning definite limiting dates to the belt as a whole. We can say only that orogenic breakup in the eastern part of the Great Basin appears to have begun in Cretaceous time, not later than the middle of that period. The growing Cretaceous highland had its eastern front remarkably near the eastern limit of thick Paleozoic deposits. In the early Cenozoic orogeny advanced eastward, and the comparatively brief life of the Mesozoic Rocky Mountain seaway was ended.

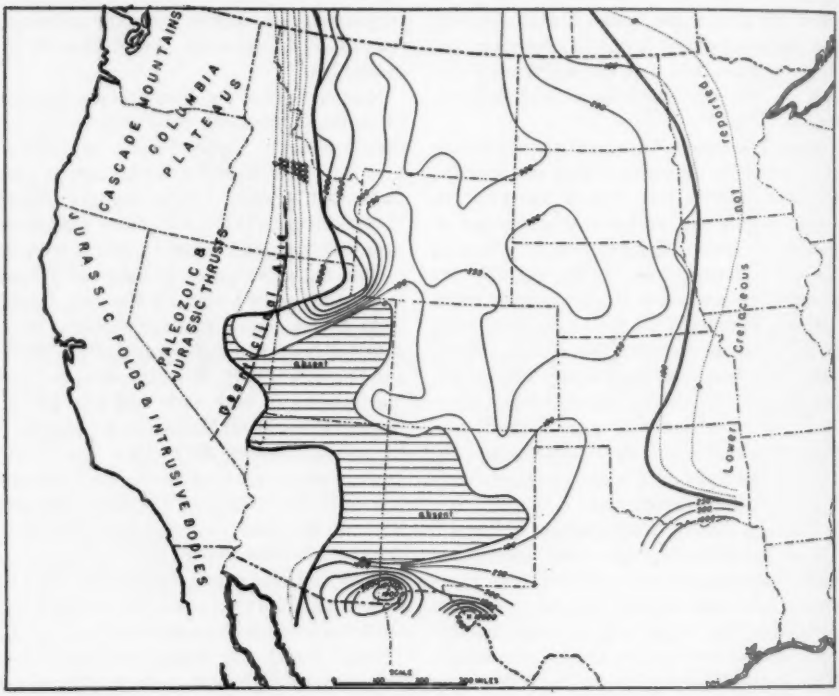


FIGURE 7.—DISTRIBUTION AND THICKNESS OF LOWER CRETACEOUS DEPOSITS IN WESTERN UNITED STATES
From Reeside (1944) with extension of map westward to show some major tectonic units.

It appears, therefore, that waves of deformation began in Jurassic time in California and western Nevada, and advanced during the remainder of the Mesozoic and part of the Cenozoic era, covering the belt, hundreds of miles wide, from the present Pacific coast to the Rocky Mountain front. Perhaps the real start should be assigned to the late Paleozoic, when large-scale deformation reached eastward to central Nevada, the Mid-Cordilleran geanticline arose farther east, and crustal unrest still farther east is reflected in the geosynclinal sediments of that time (Gilluly, 1932). However, full-dress orogeny that began in the Jurassic progressed consistently eastward. How continuous this advance was, in space and time, is not established; the lack of sedimentary record for critical intervals in much of the area is a serious handicap in reading the full story. However, as several students of the region have commented, the supposed distinction between the classical orogenic episodes long known as Nevadan and Laramide grows more uncertain as field evidence comes in, and perhaps will disappear as a partition with any real significance.

The nature of the deformation in the old geosynclinal belt holds particular interest. In nearly all areas of Paleozoic and Mesozoic rocks that have been closely studied, large-scale thrusts of low dip are a characteristic feature. On a regional map the known thrusts plotted to scale may not appear impressive (Nolan, 1943, Fig. 12). However, the thrusts as mapped are in distinct groups, each of which represents a limited area in which recent detailed field work has been done. The group in southern Nevada is outstanding, partly because the Colorado River drainage, in late geologic time, has removed large quantities of concealing basin fill, particularly under the more pluvial climate of the Pleistocene. The widely separated groups of thrusts now known emphasize again that in any generalizing about the geology of the province as a whole we have to deal with points of knowledge, separated by large areas of ignorance. However, the fact that similar groups of thrusts are found in so many of the carefully mapped areas, which have a random distribution in this wide region, is strong indication that failure on large flat thrusts has been a regional habit. Since exposure of any thrust depends on the height of the plate in relation to basin fill, and on other variables, probably many large thrusts escape notice even in areas that are carefully mapped. Some ranges whose structure appears to be a succession of close folds may be transected at depth by major thrusts.

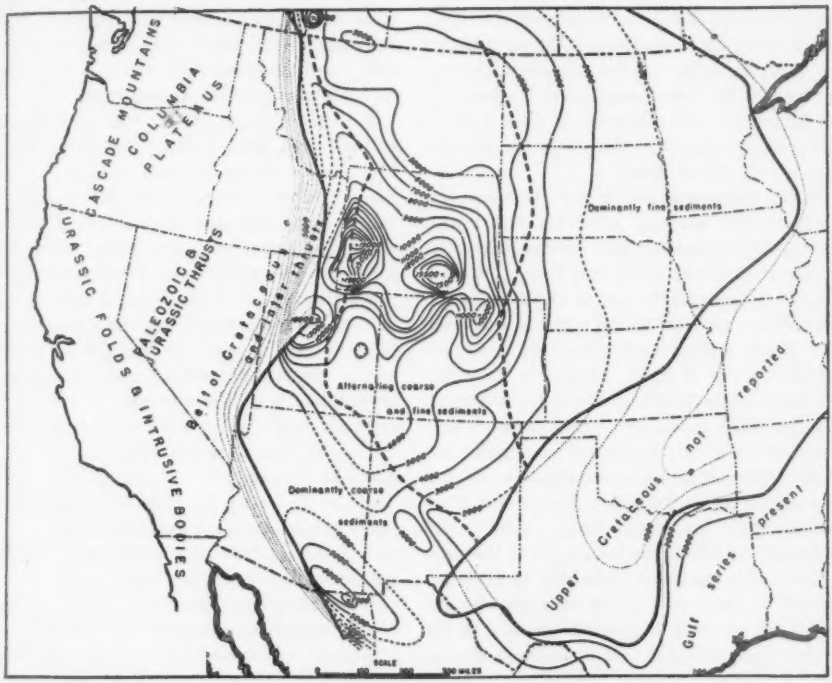


FIGURE 8.—DISTRIBUTION AND THICKNESS OF UPPER CRETACEOUS DEPOSITS IN WESTERN UNITED STATES
(From Reeside 1944 with additions as in Figure 7.)

None of the thrusts in Nevada and western Utah are known to involve the older Precambrian basement rocks. This is a striking and assuredly an important fact, in view of the deep erosion to which the deformed rocks have been subjected. The old geosynclinal section is of course very thick, and in some parts of the belt at least two and perhaps several sheets were superposed. It is possible, moreover, that older rocks in the "root" portions of some sheets are

hidden under the Cenozoic fill in wide basins. We are reminded, however, that in the Appalachian belt west of the Blue Ridge no basement rocks appear in the many imbricate thrust sheets. Thus, in both of these inner geosynclinal belts, evidence now in hand suggests that most of the thrusts and associated folds are products of comparatively shallow deformation, restricted to the sedimentary cover. Is this type of deformation characteristic of inner belts of marginal geosynclines? What was the mechanism of such deformation?

Basin Range structure, a large, complex, and controversial subject, must be dealt with briefly. Chief attention thus far has been given to the part of the Basin and Range province that lies approximately within the limits of the Cordilleran geosyncline. In much of that area the Basin Range faults trend northerly, or somewhat east of north, parallel to the structural grain established by the earlier orogenies. Farther south the eastern limit of Basin Range faulting swings southeastward far outside the area of thick Paleozoic sedimentation. In that part of the region the axes of ranges, and characteristically the Basin Range faults, trend strongly northwest, at a large angle to the common trend in the Great Basin (Fig. 1). However in this southern part of the province also the Basin Range trend is generally parallel to the earlier structural grain; for in northern Mexico and in southern Arizona the thrusts and the axes of folds developed in Mesozoic orogeny strike northwestward. Thus in general the Basin Range pattern was set by earlier tectonic events.

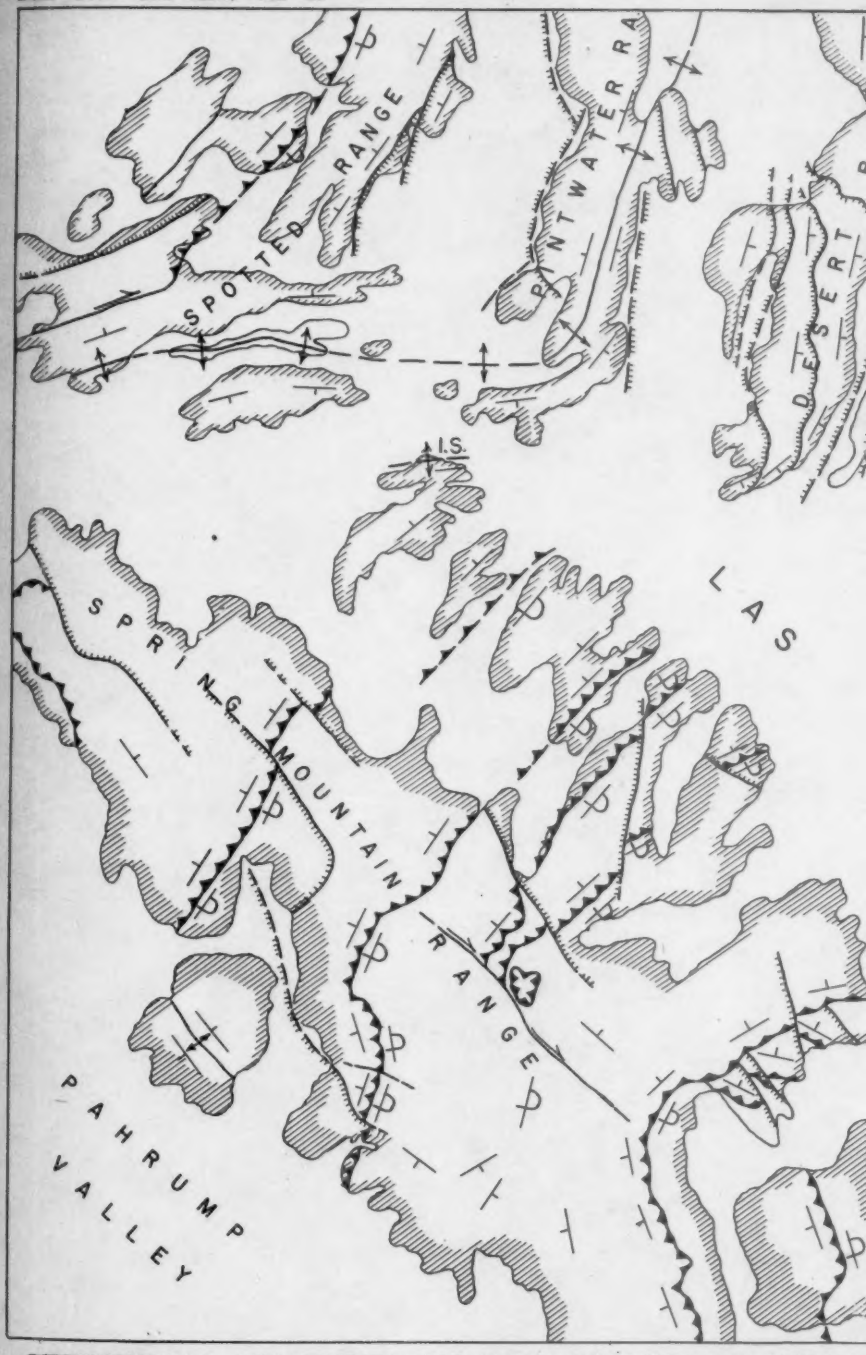
The northerly and northwesterly trends of Basin Ranges meet abruptly in southern Nevada. Clearly the junction defines an important structural zone; numerous features in this belt bear eloquent testimony to horizontal movements through a long time interval in a zone of transverse faulting, the crustal block on the south having moved relatively northwestward. The most recent record of such movement is found in the area affected by the Cedar Mountain earthquake of 1932, 200 miles northwest of Las Vegas (Gianella and Callaghan, 1934). The line of junction between ranges that trend north and those with northwest trend is remarkably parallel with and otherwise similar to the San Andreas fault zone (Fig. 1).[1] Cumulative evidence of long-continued movement in the belt of northwest-trending ranges is a stimulus to careful study in this sector of the Basin Range problem. In particular, the possible connection between the puzzling chaos structure in the Death Valley region (Noble, 1941) and movements in the belt of northwesterly trends merits investigation.

Features of the Las Vegas area appear in the tectonic map, Plate 2. Sharp bending of structure lines in the north-trending ranges reveals relative southeastward movement of a major crustal block. The axial trace of a large faulted anticline, approaching the valley from the north, swerves abruptly to a westerly course. A flat thrust west of the anticline turns into a vertical tear fault parallel to the general trend in the transverse zone. All of the blocks in this part of the area are sheared and crushed, and dolomitization of the limestone formations is extreme. At the end of each range farther east the steeply dipping strata hook sharply to the west, as do the traces of two large thrusts. South of the valley the high Spring Mountain range trends northwest, but major thrusts that extend into it from the south bend strongly northeastward toward Las Vegas Valley.

Adjacent to Las Vegas Valley, steep uptilting of the sedimentary section was incidental to movement in the zone of horizontal shear. Beds probably of Eocene age, which are tilted equally with the Paleozoic formations, were deposited after the latest thrust faulting that has been dated closely (Longwell, 1949). On this basis it may be suspected that the zone of transverse faulting developed later than the thrusting. However at least three of the thrusts are genetically connected with tear faults that are closely related to the shear zone (Pl. 2). Therefore the zone of horizontal adjustment was in operation while some of the thrusting was in progress, and continued active to a much later date.

Presumably the habitual failure by hori-

[1] This abrupt change in trend was noted by Gilbert in his early studies (1875, p. 22), and has claimed attention of later workers in the area. After Gianella and Callaghan commented on the belt marking the change in trend, Billingsley suggested for this belt the name Walker Lane (Locke, Billingsley, and Mayo, 1940, p. 523).



STRUCTURAL FEATURES IN THE RANGES ADJACENT

Traces of major thrusts are shown by black triangle
hachures on downthrown side; half-arrows show horizont
axis shown by usual convention, as are normal and overt
hidden by alluvium.

L. V., Las Vegas; I. S., Indian Springs; F. M., Frenc

OJACENT TO LAS VEGAS VALLEY, SOUTHERN NEVADA

ack triangles, with points on thrust plate; normal faults have ow horizontal movement on tear faults; traces of anticlinal al and overturned dips; broken lines indicate folds and faults

F. M., Frenchman Mountain; D. L., Dry Lake Range.

zontal shear along this zone and on the San Andreas fault zone is a response to regional north-south compressive stress. Mesozoic folding and thrusting in the Mexico-Arizona belt reflect a strong component of compression parallel to the meridian; and other tectonic units of the region, notably the Transverse Ranges of southern California and the great Uinta arch of northern Utah, point even more strongly to compression that has acted generally north-south. Relation of this force to forces responsible for north-south trends in Nevada and western Utah, and to the development of Basin Range faults in general, is a matter of large interest in tectonic theory.

In the ranges about Las Vegas the numerous large normal faults (only a few of which are shown on the map, Plate 2) are representative of such faults throughout the province. In date of movement they range from the time of thrusting, through several Cenozoic epochs, down to very recent. West of Las Vegas a wide, flat thrust plate was broken by steep faults, and the tilted blocks were overridden by a similar plate. This is not a unique example; in the Gold Hill district of western Utah, Nolan (1935) found evidence of several distinct "structural cycles" in each of which large-scale thrusting was followed by development of important normal faults. These movements began probably in the Cretaceous period, and continued or recurred through much of the Cenozoic. Near Las Vegas valley some of the north-trending normal faults developed with the steep tilting in post-Horse Spring (Eocene?), pre-Muddy Creek (Miocene?) time. Many other faults are pre-Miocene (?), others displace Miocene (?) beds, and still others cut younger deposits, including recent fans on piedmont slopes. Of several misconceptions that persist regarding Basin Range structure, none is more erroneous than the view that most of the faulting was concentrated in late Cenozoic time.

Record of Igneous Activity

It has long been recognized that Paleozoic igneous activity in western North American was restricted to a belt whose eastern margin was only a moderate distance east of the present Pacific coast (Lindgren, 1915). Rocks that provide this record are predominantly volcanic, largely andesitic, but in part of more mafic composition. In the Sierra Nevada area, masses of diorite were intruded probably in late Paleozoic time (Knopf, 1929), and in British Columbia widespread granitic injections were emplaced sometime after the Lower Cambrian and possibly before the Pennsylvanian (Armstrong, 1949). In Permian time the belt of extensive volcanism reached eastward into northwestern Nevada (Nolan, 1943, p. 157) and central Idaho[2] (Ross, 1934); probably this activity was connected with the late Paleozoic orogeny in its eastward advance.

Triassic volcanic rocks are abundant as far east as southwestern Nevada (Muller and Ferguson, 1936), and bentonitic clays in the Upper Triassic section of central Arizona (Allen, 1930) may represent air-borne ash from a source in Nevada or California. Jurassic deposits known in west-central Nevada, representing only the lower part of the system (Muller and Ferguson, 1936), contain great quantities of volcanic material; doubtless younger formations of this system, if they were present in that part of the region, also would record intense igneous activity. In the eastern part of the province and farther east, no igneous rocks known to be of Jurassic age are recorded. However, lavas and pyroclastic materials are included in Lower Cretaceous deposits as far east as southwestern New Mexico (Lasky, 1947). The lowest Upper Cretaceous strata of southern Nevada contain conspicuous quantities of volcanic ash (Longwell, 1949), and probably explosive volcanoes were numerous in the Cretaceous highland west of the Rocky Mountain geosyncline, supplying widespread lavas as well as pyroclastics, including the ash from which developed extensive sheets of bentonite that are common in the Upper Cretaceous section as far east as Kansas, Nebraska, and the Dakotas. Important centers of volcanism appeared within the geosynclinal belt itself in Eocene and later Cenozoic time, during and following deformation by folding and thrusting.

Thus, although it is not possible at present to assign exact dates to many important tectonic features and to large bodies of volcanic rock in

[2] The Permian date of the volcanic rocks in Idaho is not firmly established.

the Cordilleran region, in at least a general way the volcanic activity advanced eastward with orogeny. This is true also of the Mesozoic and Cenozoic plutonic history. Lindgren inferred from the field evidence that "the batholithic intrusions of the Pacific coast gradually spread eastward" (1915, p. 260). There is general agreement that intrusive bodies exposed in the western part of the Great Basin are related to those of the Sierra Nevada, and are definitely older than plutonic masses (later than Precambrian) in the Southern and Middle Rockies. According to one view, graphically expressed on the geologic map of the United States, all of these masses from the coastal belt eastward belong in two definite age groups, separated by an irregular north-south line through central Nevada; on the west are the bodies classed as late Jurassic or early Cretaceous ("Nevadan"), on the east the Tertiary intrusive bodies ("Laramide"). With improved dating of volcanic rocks in the region it appears increasingly plausible that the plutonic bodies may represent a large part of the interval from late Jurassic to Tertiary, with progressive decrease in age eastward. Such a sequence, if it should be established, would be consistent with what is now known about the regional tectonic history.

Volcanic activity in the Basin and Range province was recurrent or continuous throughout Cenozoic time. Resulting lavas and other products range widely in composition, from rhyolitic to basaltic, with repeated variations at many localities. This activity probably is related to the crustal movements responsible for Basin Range structure.

Larger Implications and Problems

Let us now examine briefly the larger framework. A relief map of the continent shows that the Basin and Range province lies in the widest segment of the North American Cordillera. The mountain belt as a whole is conspicuously parallel to the margin of the continent; and neglecting for the moment the outer Cenozoic zone, this major unit testifies, through thousands of miles of its length, to orogenic movement that advanced from the coastal belt eastward. The broad picture of this advance is represented in a series of maps published by Eardley (1949).

In recent years an important focus in studies of orogeny has been on island arcs. In a general view the American side of the Pacific is in sharp contrast with the opposite side, where a vast network of island chains extends from New Zealand northward past Asia, finally connecting with Alaska through the great Aleutian arc. In that complex mesh of curved units the general rule is convexity outward from the lands toward the Pacific. Students of Asiatic tectonics, from Suess (1904–1909) to Kobayashi (1941), state categorically that orogenic development has been and still is *centripetal* to the Pacific. The Japanese Islands have grown from northwest to southeast—that is, from the inner side outward—by successive deformation of geosynclines which have developed consistently on the outer side of the arc. Neighboring arcs are in various stages of development (Hess, 1948). Large thrusts dipping landward, and folds overturned toward the ocean, are charcateristic features.

Has there been a fundamental difference in tectonic behavior on the two sides of the Pacific? Or do we witness merely different stages in developments which in the long view may be similar? At present there are no island arcs along the entire western border of the two American continents, from southern Alaska to Cape Horn. Instead, generally deep water approaches close to the continental margins, relieved only by scattered islands, including small groups of volcanic origin; and inland the Cordilleras stretch, with few breaks, the length of the two continents.

Were there, in past eras, island arcs off western America similar to those now in the western Pacific? There was land of some kind near the present west coast of Canada and the United States, as attested by the sedimentary record found in several geologic systems, from early Paleozoic to Jurassic. In the islands of southeastern Alaska thick sedimentary sections, repsenting several Paleozoic systems, contain an abundance of boulder conglomerate, graywacke, and other clastics (Buddington and Chapin, 1929). The Franciscan deposits of California were, according to Taliaferro's interpretation (1943), derived from a Jurassic land that lay west of the present coast. Eardley (1947) has offered the hypothesis that volcanic archi-

pelagoes, including large islands such as those of modern Japan, existed in our coastal belt, with a long history of sedimentary deposition, orogeny, and igneous activity. His hypothesis carries the picture of great island arcs, convex to the Pacific, one of which formed the western rim of the wide Paleozoic seaway that covered the area of the Great Basin. If we entertain this concept, with the assumption that the vanished island arcs were closely analogous to the modern Asiatic archipelagoes, certain implications are suggested, as follows: (1) Island-arc development includes a long period of incubation and growth; for the supposed arcs west of this continent the period occupied a large part of Paleozoic and early Mesozoic time. (2) The final chapter in the history of these supposed arcs coincided with a great surge of orogeny that not only deformed the sedimentary accumulations inside the arcs, but carried far beyond, into the continental mass. (3) If we are to assume that the earlier growth of these arcs was toward the Pacific, as is said to be true of the Asiatic examples, then in late Paleozoic time the direction of orogenic advance on the American side must have been reversed. Arcs that may have lain in the coastal belt were left far in the rear as the orogenic front rolled eastward.

Paleozoic deformation that reached far into the Great Basin was not accompanied by intrusion of igneous masses that have become exposed. Jurassic deformation, with development of the great batholiths of the coastal belt, marked the first clear step in the eastward march. This was a diastrophic event of the first magnitude. Although folding and thrusting in the Sierra Nevada and the Coast Range of Canada does not appear to match Alpine deformation in severity, the plutonic masses formed in this belt are unsurpassed in linear extent. From Alaska to Lower California the length of the strip in which these bodies are exposed exceeds 2500 miles. This stupendous mass of siliceous magma (if it was magma!) may have been the product of mysterious plutonic processes during long Paleozoic periods while volcanoes in the coastal belt were discharging lavas and pyroclastics. The intense and prolonged burst of activity in the Triassic and early Jurassic may have been a signal that the brew so long in preparation was nearly ready to serve.

Recently we have heard in brief outline two versions of a tectonic concept called by the authors the "blister hypothesis" (Rich, 1937; Wolfe, 1948). The Coast Range-Sierra Nevada chain of batholiths may be suggested as the summit of a gigantic *blister*. One author provides, as an important part of the blister mechanism, gravitational sliding on the flanks of the uplift, with resultant large-scale folding and thrusting. The flat thrusts and related folds of the Great Basin present a problem of origin, and the sliding hypothesis is appealing. In particular, the Jurassic deformation in western Nevada appears to be favorably situated in relation to the initial Sierra Nevada uplift. But unfortunately for this form of the sliding concept, some of the most intense deformation of that western belt occurred in early Jurassic time, before the Sierra batholith was emplaced. Moreover, this blister must have had a western as well as an eastern flank, but no orogeny closely following the batholithic intrusion is reported on the west.

The belt of Cretaceous and later thrusts in the eastern part of the Great Basin lies mainly east of the Mid-Cordilleran geanticline (Pl. 3). Perhaps this uplift, started in Permian time, became an elongate blister early in the Cretaceous period. Indeed, the Idaho batholith appears to lie in the course of the uplift, and this plutonic mass may be of Cretaceous date, since field evidence now known makes it merely post-Triassic. But although the aggregate displacement on thrusts in eastern Nevada and western Utah amounts to tens of miles, no clue to a contemporaneous blister is found anywhere in that part of the region, to mark the summit from which sliding may have proceeded. In particular, that part of the province has not been flooded with basaltic lavas, as is postulated for the belt at the crest of a riven dome.

Probably the doming envisaged by proponents of the blister mechanism is of larger scale than that suggested above. Rich (1937) speaks of "sliding of the crust above the molten interior," following upward expansion in a large segment affected by radioactive heating. The downward component of weight in the sliding mass "would initiate geosynclinal downbending,

which would be increased by the weight of sediments." Finally the crust would shear across at the edge of the dome, piling up "in mountainous thrust sheets, crumpling the sediments of the inner margin of the geosyncline." Major events in Cordilleran history do not appear to fit logically into this scheme. The Rocky Mountain geosyncline was initiated in the late Jurassic, approximately at the time major deformation occurred in the Sierra Nevada-Klamath belt. Are these events chargeable to sliding on opposite flanks of a blister-dome in the Great Basin area? If so, there should have been contemporaneous tensional failure at the crest of the dome, with block faulting and effusion of plateau basalts. How could such effects be postponed until Cenozoic time?

If we assume two blister-domes, one west of the present coast as the cause of Nevadan orogeny, another in the Great Basin area to explain Laramide development, spatial and temporal inconsistencies still confront us. The assumed mechanism must account for: (1) the wide belt of major thrusts, at least in part of Cretaceous date, on the site of the Cordilleran geosyncline; (2) the formidable mountain land that shed stupendous quantities of clastic sediments eastward into the Rocky Mountain geosyncline, and westward into a Pacific trough; (3) eventual deformation of the Rocky Mountain geosyncline, not merely at its western margin but through hundreds of miles of width, while the coastal trough remained undisturbed. Moreover, the one massive effusion of basaltic lavas in the region occurred far from any logical position of a domal crest, and largely during Miocene time, two geologic epochs later than the orogeny charged to assumed crustal sliding. The scale and versatility demanded of a blister-dome that is to be held accountable for such colossal and varied feats seem hardly credible.

The blister concept is of course strongly reminiscent of Haarmann's oscillation theory (1930), which assumes sliding on the flanks of migratory "geotumors." If a reasonable case could be made for the origin and the supposed behavior of "geotumors," the mechanism provided by them might in part meet requirements set by the known facts in Cordilleran tectonic history. Outstanding among these facts are the eastward advance of orogeny across the Cordilleran geosyncline, development of the Rocky Mountain geosyncline while mountains grew at its western border, deformation of the younger geosyncline in its turn, and structural collapse of the highlands to the west. Even if Haarmann's scheme might appear plausible in connection with this chain of events, however, we look in vain for evidence of a "geotumor" that may have led to yet another and much more recent development—the renewal of orogeny along the present coastal belt in late Cenozoic time. This failure in an apparently elementary test is a serious blow to the Haarmann concept as an acceptable diagnosis of the orogenic process.

Development of the Rocky Mountain geosyncline holds commanding interest among major events of Cordilleran history. In comparison with other great belts of deposition that have been deformed, the life of this one was brief and dramatic. Starting in quiet Jurassic seaways east of the Mid-Cordilleran geanticline, deposition and attendant subsidence were renewed in the Cretaceous period, and quickened with mounting deformation that advanced from the west (Spieker, 1949). Development of the geosyncline may reasonably be considered incidental to the eastward surge of orogeny. Its life span—approximately one geologic period—contrasts sharply with the nine periods represented by thick deposits in the Cordilleran geosyncline. Rate of accumulation in the Rocky Mountain belt averaged much faster than in the older trough, if we consider the complete record in each. However, total thicknesses of strata in the Cretaceous belt that reach 20,000 feet are exceptional, and sections with half this thickness are more common. These facts do not bear out a view expressed by some students that 7 or 8 miles of geosynclinal subsidence has been required to initiate final deformation.

Schuchert (1923) described the Rocky Mountain geosyncline as *sequent*, because in a considerable part of its length it is superposed on the earlier Cordilleran belt of deposition. The adjective sequent is not appropriate in the United States, where the younger trough chose a course east of the older feature and in part on the site of late Paleozoic highlands (Lovering, 1932). Thus the Rocky Mountain geosyncline has several peculiarities, but in classifi-

cations now in vogue it is not cited as a type, with a unique descriptive term. Let us hope it may escape a cumbersome designation, additional to the unwieldly lists now at our disposal.

Structure in the Rocky Mountain-Colorado Plateaus belt is in sharp contrast with that in the Great Basin. The enormous Plateaus block has been warped, locally buckled, broken by steep faults; but in the main, except for large uplift, it has come through the regional orogeny with surprisingly moderate change. East and north of the Plateaus the Rocky Mountain structure is dominated by great archlike uplifts and upthrust blocks, the eroded crests of which expose Precambrian basement rocks over wide areas (Pl. 3). There are important thrusts and belts of folded sedimentary rocks, but faulted arches of large span and amplitude are an outstanding feature. This structure seems to indicate deformation irregularly distributed and extending to great depth, affecting a thick shell. The belt of uplifted basement rocks is in the form of a wide arc, convex to the east, directly opposite the Basin and Range area, as if there were some genetic relationship. Farther north, in northern Montana and Canada, the Rocky Mountain foreland has no such uplifts exposing the basement; major thrusts and folds there bring the Beltian and Paleozoic formations up to form an abrupt and comparatively regular front overlooking the weaker formations of the plains.

Thus the northern Basin and Range region, which is an exceptional segment of the interior Cordilleran belt, has directly in front of it an exceptional segment of the Rocky Mountains. The Cretaceous highland of the Great Basin area, which supplied immense quantities of sediment to the Rocky Mountain geosyncline east of it, has not behaved in its later history as many other old mountain belts have done. No axial element of this former highland can be identified at present; and the belt has not risen by recurrent upwarping to mark successive cycles of erosion. Instead, generally in unison with deformation and uplift of the Rocky Mountain geosynclinal area, there has been progressive breakup and collapse of the old highland mass. It is as if material at depth had been transferred from the Basin and Range belt to the Rocky Mountain country. In this suggestion we of course enter the realm of frank speculation. However, we grope for any logical suggestion that may explain the tectonic development of the region, having regard for the principle of isostatic balance. A common assumption has been that in an orogenic zone the sialic shell has been thickened by folding and thrusting in bedded rocks, accompanied by intimate shearing of the massive basement rocks beneath. The peculiar structure of the Southern and Middle Rockies does not seem to accord in simple fashion with this concept. Moreover the great Colorado Plateaus block is an important unit in the wide belt that has been transformed from sea floor to mountain land. By what mechanism may the low-density shell in this unit of simple structure have been thickened to give it equilibrium at its greatly increased altitude? The same general problem is, of course, presented by the Altiplano of the Andean highland, the Thibetan Plateau north of the Himalaya, and similar units in other major mountain belts.

Let us therefore entertain the hypothesis that slow transfer at depth, in what Albrecht Penck called the "geoplasmatic region," was a significant mechanism in the rise of the Rocky Mountains and Colorado Plateaus, and the generally concurrent breakup within the Basin and Range province. This hypothesis accords with Gilbert's suggestion, which he offered in bare outline (1875, p. 62), that Basin Range structure may be explained as adjustment within an outer passive shell to deformation in a deeper zone. The nature of this deep-seated action he did not attempt to define. We may speculate that it was plastic deformation, accompanied by actual transfer of rock material from west to east. Such a mechanism, whatever the driving force, may best explain the final disappearance of land that existed west of our present Pacific coast; witness the irregular submarine topography off southern California, which as Shepard points out has a striking resemblance to some areas in the Basin and Range province (Shepard and Emery, 1941). Moreover the land, whether a unit mass or an archipelago, that lay east of the Appalachian geosyncline vanished mysteriously after orogeny ended. What we know of Triassic faulting along the Appalachian belt, together with seismic evidence that basement rock is now at great depth under the shelf,

accords with the concept that collapse of the old land occurred as the final wave of orogeny carried westward.

Possible implications of this hypothesis are far reaching. If there is a mechanism whereby masses of low-density rock are slowly transferred at depth, large-scale fragmentation and foundering of continental masses may have occurred, in accord with the principle of isostatic balance.

The thought of deep-seated landward flowage across the continental margins brings to mind Lawson's concept (1948) of such movement in response to secular erosion that lightens the lands and weights the ocean floors. Surely this surficial transfer of load must be an important quantity in the equation of crustal deformation. However, as Lawson's analysis makes clear, we cannot accept the isostatic mechanism, working alone, as sufficient to renew the potential required for the age-long recurrence of orogeny; to do so would be indorsement of "bootstrap mechanics." Moreover, those who urge isostatic flow as a primary agent in orogeny must face the sobering fact that great stretches of continental borders, particularly those facing the Atlantic, have long been orogenically quiescent. In league with other forces, isostatic adjustment may be a powerful contributory factor in localizing and perhaps hastening deformation.

The primary agent of crustal deformation—that is the supreme mystery. Advocates of convection currents in the body of the earth will urge this concept as the most promising to explain the slow but insistent eastward march of Cordilleran orogeny; this suggestion seems as good as any yet offered in this no-man's-land of geologic thought. The choice among hopeful solutions of the orogenic riddle has not narrowed—within recent decades it has grown considerably more varied, a sure indication that the field remains nebulous. Near the beginning of my teaching experience an older colleague once asked which of the hypotheses of orogeny I favored. On my profession of a mind entirely open, unpledged to any creed, he said with paternal firmness, "Young man, you must learn to make up your mind in these fundamental matters, and let people know just where you stand." In spite of this friendly admonition and the passage of years, my mind is not yet made up; but I do not find that fact a cause for unhappiness. Like the Swiss geologists who toiled through two generations on the puzzling complex of the pre-Alps, we must be reconciled not to know the basic answers, while we devote our interest and our energies to finding out *exactly what is to be explained*. To that end I look with covetous eyes on the vast store of information, much of it doubtless of critical value, that is still to be harvested in the Basin Ranges.

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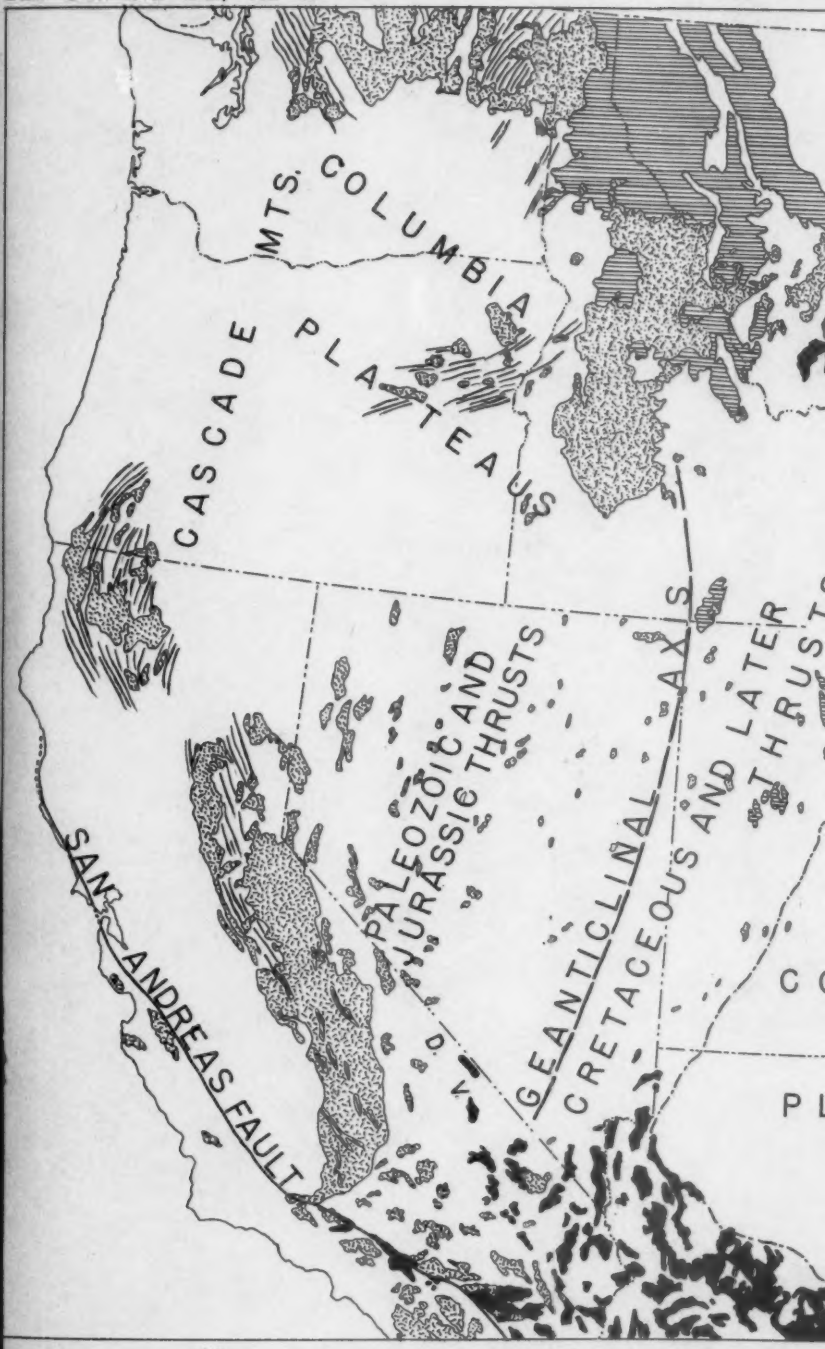

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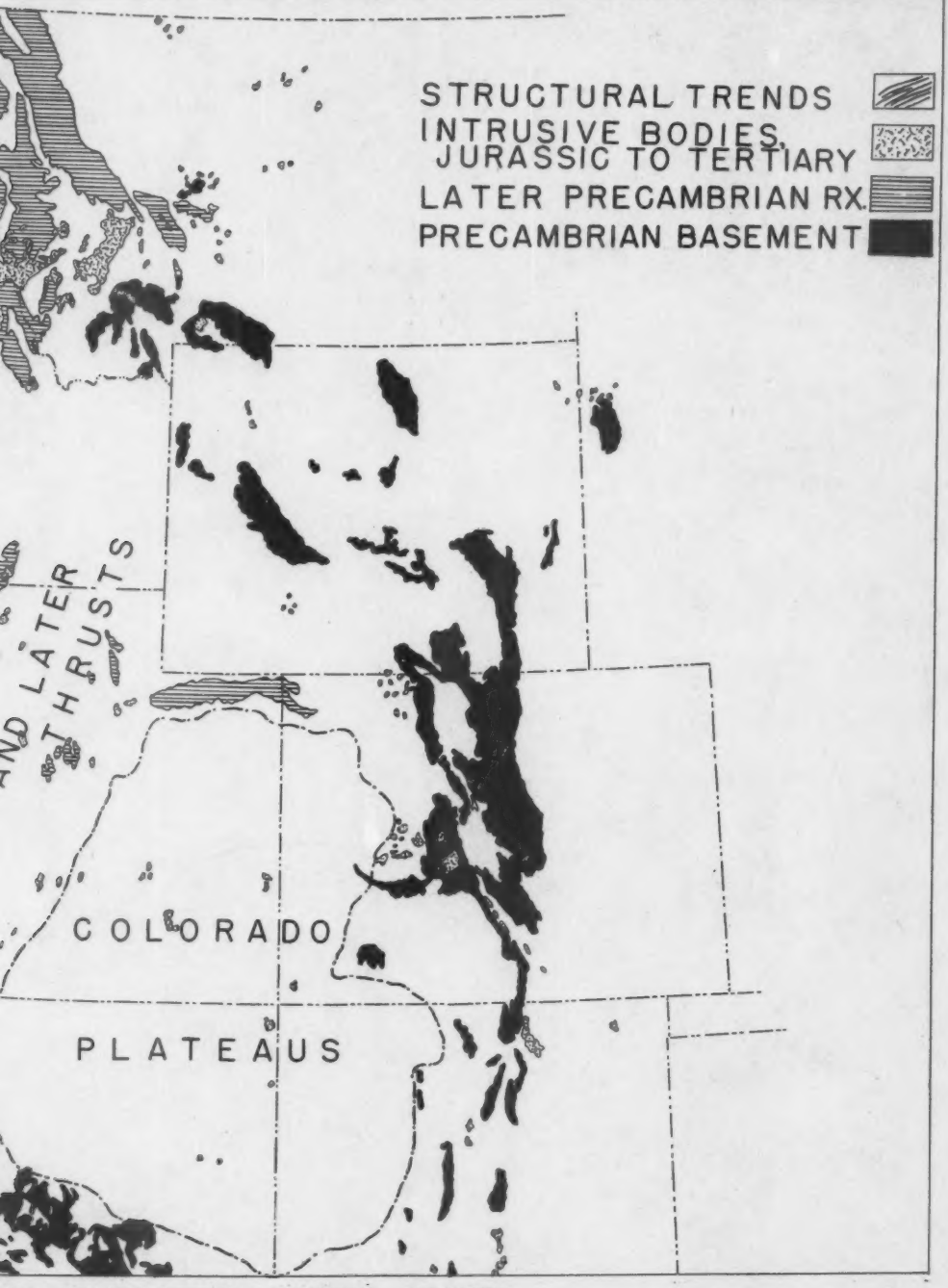

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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STRATIGRAPHY AND THE STUDY OF METAMORPHIC ROCKS

BY MARLAND P. BILLINGS

(Address of Retiring Chairman, Section E, American Association for the Advancement of Science, New York, December 1949)

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ABSTRACT

In recent years many students of metamorphic rocks have become so preoccupied with minor structures, structural petrology, physical chemistry, and granitization that the stratigraphy of metamorphic rocks has been neglected. There is great danger that the younger men, indoctrinated with the idea that stratigraphy and sedimentation are unrelated to metamorphic geology, will be inadequately trained to study metamorphic rocks in the field. Geologic maps of regions characterized by different grades of metamorphism should be based on stratigraphy. The assignment of the rocks to formations should be based on the inferred lithology prior to metamorphism and should not be based directly on the present lithology. The student of metamorphic rocks should be familiar with modern concepts of stratigraphy and sedimentation, such as deposition in transgressing and regressing seas, changes in sedimentary facies, and time surfaces. Metamorphic geologists should be familiar with the textures, mineralogy, and chemical composition of sedimentary rocks. When sufficiently large areas are studied, the metamorphic geologist should think in terms of the paleogeography and climatic conditions at the time of sedimentation. Stratigraphers and students of historical geology should realize that their concepts of paleogeography will be incorrect if they neglect the wealth of data to be obtained from metamorphic rocks. These facts can be abstracted from the metamorphic rocks only by intensive investigations in the field by a host of metamorphic geologists well-trained in stratigraphy.

INTRODUCTION

During the last few decades several excellent books dealing with metamorphic rocks and metamorphism have been published. Among others may be mentioned those by Grubenmann and Niggli (1924), Harker (1932), Grout (1932), Eskola (1939), and Turner (1948). Discussing closely related subjects are the books by Knopf and Ingerson (1938), Fairbairn (1942), and E. Cloos (1946). They are all fine treatises on metamorphism and related mat-

ters; obviously all students of metamorphic rocks should be masters of the contents of these books. Nevertheless, paradoxical as it may first sound, a geologist who has completely assimilated the principles presented in these publications may be inadequately trained to study metamorphic rocks in the field. He may be a good mineralogist and petrographer, he may be familiar with what is known about the physical chemistry of metamorphic rocks, he may know structural petrology, and he may be familiar with lineation and other minor structures. But he may lack training in stratigraphy. Within the field of stratigraphy I include sedimentation and paleogeography as well as stratigraphy in the narrower sense.

Some of my listeners may say that it is axiomatic that the metamorphic geologist should be well trained in stratigraphy. They could cite many cases of excellent stratigraphic work done in metamorphic rocks, as, for example, the studies by Noble and Harder (1948) in the Black Hills of South Dakota. Perhaps to these metamorphic geologists, cognizant of the importance of stratigraphy, these remarks will appear unnecessary. But over a period of years it has gradually become apparent to me that many geologists do not realize the importance of stratigraphy in the study of metamorphic rocks. Moreover, geology, particularly metamorphic geology, structural geology, historical geology, stratigraphy, and paleogeography, have suffered because of this failure.

Several pieces of evidence may be cited to show that the significance of the application of stratigraphic principles to the investigation of metamorphic rocks has not been fully appreciated and that the stratigraphic approach to metamorphic geology is becoming more and more neglected by some groups. I have always realized that many students of metamorphism, including some of the more prolific writers, have been so preoccupied with petrography, physical chemistry, structural petrology, and granitization that they have neglected stratigraphy and sedimentation. I have also discovered that many graduate students, both at my own university and elsewhere, have a warped attitude toward the study of metamorphic rocks. They think that petrography, physical chemistry, structural geology, and structural petrology will give answers to problems that can be solved only by stratigraphy. I do not mean to discredit these other fields. For example, the study of metamorphic rocks received a tremendous impetus from the application of the principles of physical chemistry (Goldschmidt, 1912; Johnston and Niggli, 1913). I merely regret that stratigraphy has been neglected.

But the principal instigation of the presentation of this paper was a remark a stratigrapher made to me a year ago. I was attending the *Symposium on Sedimentary Facies in Geologic History* held at the meeting of the Geological Society of America in New York in November, 1948 (Longwell, 1949). When my friend, the stratigrapher, suddenly saw me, he tapped me on the shoulder, and in a voice that combined surprise and amusement asked simply: "What are you doing here?" That simple query expresses most eloquently the necessity of these remarks. It indicates only too clearly the great gap that exists between stratigraphy and metamorphism in the minds of some geologists; they think that "soft-rock geology" and "hard-rock geology" are two radically different fields.

Metamorphic Zones or Facies

I hope that any stratigrapher who has been inveigled into listening to me because of my title will bear with me for five minutes while I discuss briefly, and perhaps inadequately, the subject of metamorphic zoning or metamorphic facies. I shall then return to terminology and concepts with which stratigraphers are more familiar.

Metamorphism involves recrystallization of rocks and the development of secondary structures. In the process of recrystallization, the existing minerals may merely change in size or shape, as when a limestone changes into a marble by the formation of larger but fewer calcite grains. But recrystallization often involves the formation of new minerals, which is necessarily accompanied by the destruction of some of the older minerals; some petrographers refer to this type of recrystallization as neomineralization. The secondary structure that may develop during metamorphism is variously called foliation, cleavage, or schistosity. Recrystallization without the formation of a sec-

ondary structure is possible in contact metamorphism; and even in regional metamorphism a final recrystallization can destroy whatever secondary structures may have previously existed. Conversely, a secondary structure can form without recrystallization or at least without the formation of new minerals; fracture cleavage is an example of such a structure.

During the last few decades the principle of metamorphic zoning or metamorphic facies has become well established. The zones or facies are based on the mineralogy of those rocks most sensitive to progressive changes during metamorphism; argillaceous sediments and basaltic rocks are especially susceptible to change. The intensity or perfection of the secondary structures is not taken into consideration in assigning a rock to the appropriate zone or facies. I emphasize this last point because more than once I have heard a stratigrapher say that a slate with a strong cleavage diagonal to the bedding was highly metamorphosed. Perhaps from the point of view of the seeker of fossils such a rock is hopelessly far gone and hence highly metamorphosed, but to the student of metamorphic rocks a slate is an example of low-grade metamorphism.

In the time available it is obviously impossible to give a thorough review or analysis of the several zoning or facies classifications employed by different students of metamorphic rocks. The differences are more apparent than real, inasmuch as all the classifications are based on the mineralogical associations in the rocks.

The isograd system initiated by Barrow fifty years ago may be described first, and then the other systems may be briefly compared with it (Turner, 1948, p. 35). Barrow recognized that in going from an area of unmetamorphosed sedimentary rocks into progressively more highly metamorphosed rocks new minerals appear in orderly succession. Thus, in a series of argillaceous rocks subjected to regional metamorphism, the first index mineral to appear is chlorite, followed successively by biotite, garnet (almandite), staurolite, and sillimanite. A line can be drawn on the map indicating where biotite first appears. This line is the biotite isograd; the less metamorphosed argillaceous rocks on one side of this line lack biotite, whereas the more metamorphosed rocks on the other side contain biotite. An isograd can be drawn for each new mineral. In Figure 1 the isograds are shown as dotted lines and are appropriately labeled by capital letters: *B* for the biotite isograd, *G* for the garnet isograd, *St* for the staurolite isograd, and *Si* for the sillimanite isograd.

Actually the isograds are surfaces, and the lines we draw on the map are the intersections of these surfaces with the surface of the earth. With the increase in metamorphism there is generally a progressive increase in the grain size of the matrix of the rocks, although some porphyroblasts, such as staurolite, may be exceptionally large. Clastic texture also tends to disappear with the increase in metamorphism, and pebbles in conglomeratic rocks are difficult or impossible to recognize in the more highly metamorphosed rocks.

Argillaceous rocks containing metamorphic chlorite but none of the other index minerals are said to lie in the chlorite zone; rocks between the biotite and garnet isograds lie in the biotite zone; those between the garnet and staurolite isograds are in the garnet zone; those between the staurolite and sillimanite isograds are in the staurolite zone; and those rocks containing sillimanite are in the sillimanite zone. I should emphasize again that this scheme applies directly only to argillaceous rocks, and some of these index minerals will appear only in such rocks. A pure limestone, composed only of calcium carbonate, generally shows no change in mineralogy throughout all these zones. However, argillaceous rocks are present in most stratigraphic columns, and consequently a zoning classification based on them is readily employed in most areas.

The chlorite zone is sometimes referred to as the low-grade zone; the biotite, garnet, and staurolite zones are collectively called the middle-grade zone; and the sillimanite zone is called the high-grade zone (Billings, 1937).

To describe the zonal classification of Grubenmann and Niggli (1924) and the facies classification of Eskola (1939), and especially to consider the relative merits of all the classifications, would lead into such a long digression that the main purpose of the present paper would be-

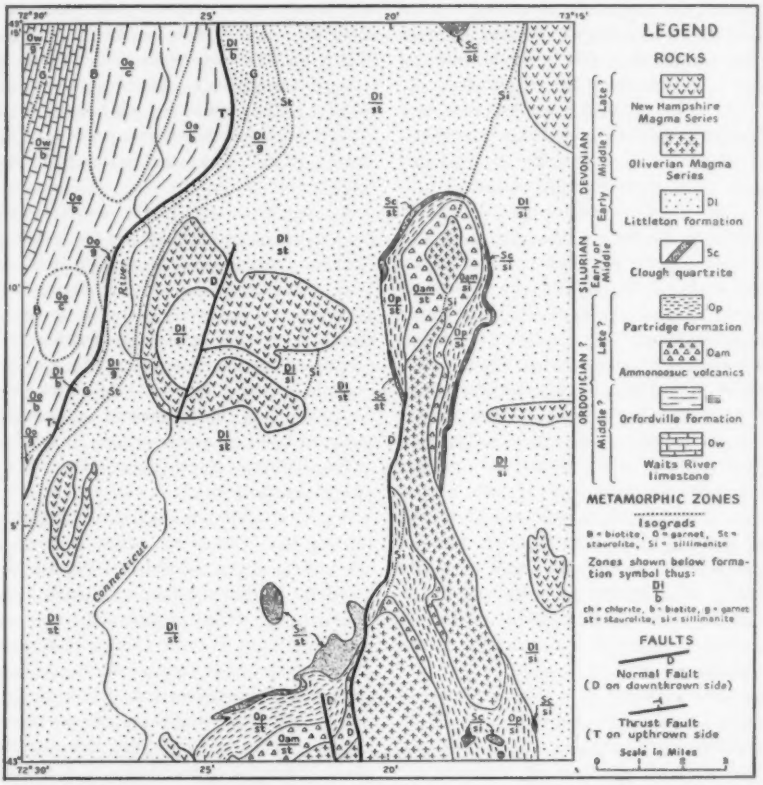

FIGURE 1.—CARTOGRAPHIC REPRESENTATION OF STRATIGRAPHY AND LITHOLOGY IN AREAS OF PROGRESSIVE METAMORPHISM

Geologic map of Bellows Falls quadrangle, New Hampshire-Vermont, after F. C. Kruger (1946). *Waits River formation* (*Ow*)—Zone *b:* phyllite and calcareous phyllite, both of which may contain porphyroblasts of biotite; also impure marble. Zone *g:* phyllite and calcareous phyllite, both of which may contain porphyroblasts of biotite and/or garnet; also impure marble. *Orfordville formation* (*Oo*)—Zone *c:* slate, black phyllite, quartzite, and chlorite schist. Zone *b:* gray phyllite, black phyllite, quartzite, all three of which may contain porphyroblasts of biotite; also albite-epidote amphibolite. Zone *g:* gray phyllite and black phyllite, both of which may contain porphyroblasts of biotite and/or garnet. *Ammonoosuc volcanics* (*Oam*)—Zones *st* and *si:* amphibolite and some biotite gneiss. *Partridge formation* (*Op*)—Zone *st:* mica schist, mica-quartz schist, and impure quartzite, all three of which may contain porphyroblasts of biotite, garnet, and/or staurolite. Zone *si:* coarse mica schist, mica-quartz schist, and micaceous quartzite, all three of which may contain porphyroblasts of garnet and/or sillimanite. *Clough quartzite* (*Sc*)—Zone *st* and *si:* quartz conglomerate, quartzite, and quartz-mica schist. *Littleton formation* (*Dl*)—Zone *b:* phyllite, quartz phyllite, and quartzite, all three of which may contain porphyroblasts of biotite. Zone *g:* phyllite and quartz phyllite, both of which may contain porphyroblasts of biotite and/or garnet. Zone *st:* mica schist, mica-quartz schist, and quartz-mica schist, all three of which may contain porphyroblasts of biotite, garnet, and/or staurolite. Zone *si:* coarse mica schist, sillimanite schist, mica-quartz schist, quartz-mica schist, and quartzite.

come concealed. It will suffice to say that the rocks in the epizone of the Grubenmann-Niggli classification are similar to those in the chlorite zone of the isograd classification; those in the mesozone are similar to those in the biotite, garnet, and staurolite zones; and those in the katazone are similar to those in the sillimanite zone. The rocks in the green schist facies of Eskola's classification are similar to those in the chlorite and biotite zones; those in the albite-epidote amphibolite facies are similar to those in the garnet zone; and those in the amphibolite facies are similar to those in the staurolite and sillimanite zones.

Cartographic Representation of Stratigraphy and Metamorphic Zones

The changes in metamorphic grade or facies introduce a difficult cartographic problem for the geologist trying to show the stratigraphy. In traversing a belt a few miles wide a formation that was originally shale may appear in successive zones as slate, biotite phyllite, biotite-garnet phyllite, biotite-garnet-staurolite schist, and biotite-garnet-sillimanite schist. The mineralogy of many other rocks, such as arenaceous dolomites or basalts, will show corresponding changes.

The metamorphic geologist is at once faced with a dilemma. Will his formations be based on present lithology? If so, the same stratigraphic unit would be mapped as half a dozen different formations. Such a policy would completely obscure the stratigraphic relations and would effectively conceal the data essential to unravel the geological history. The only alternative is to think in terms of the original sedimentary or volcanic formations before metamorphism. Thus a formation may be slate in one part of the area, biotite-garnet-staurolite schist in a second place, and biotite-garnet-sillimanite schist in a third place. The lithologic character can be shown by a subscript placed beneath the letter symbol that indicates the formation. Thus, adopting a system we have used in New Hampshire, on colored geological maps the same color pattern is used for a formation regardless of the metamorphic zone.

Figure 1, based on maps of the Bellows Falls quadrangle in New Hampshire and Vermont (Kruger, 1946), illustrates the cartographic method that is advocated. The metamorphosed sedimentary rocks range in age from Middle Ordovician (?) to Lower Devonian. Each formation is shown by a single distinctive pattern, the same pattern being used regardless of the degree of metamorphism. Moreover, the upper part of the bipartite letter symbols used on the map indicates the system and formation in the usual way. For example, *Dl* means Devonian (*D*) Littleton formation (*l*). The metamorphic zone is shown by the lower part of the bipartite letter symbol, *c* for the chlorite zone, *b* for the biotite zone, *g* for the garnet zone, *st* for the staurolite zone, and *si* for the sillimanite zone.

The lithology of each formation must be separately listed for each zone in which the formation appears. On a colored geological map the description of the lithology in the successive zones can easily be set up in type in the legend at the side of the map. Examples are shown in papers by Billings (1937), Hadley (1942), and G. E. Moore (1949). In Figure 1, partly to save drafting expenses, the lithology is indicated in the caption beneath the map. The symbol *Dl/si* means the Devonian Littleton formation in the sillimanite zone. The caption shows that in this zone the Littleton formation consists of coarse mica schist, sillimanite schist, mica-quartz schist, quartz-mica schist, and quartzite. The symbol *Dl/b* means Devonian Littleton formation in the biotite zone, where the rocks are phyllite, quartz phyllite, and quartzite, all three of which may contain biotite porphyroblasts.

For those preferring the facies classification such subscripts as *g*, *ae*, and *am* could be used to indicate, respectively, the green schist, albite-epidote amphibolite, and amphibolite facies.

The advantage of such a cartographic system is apparent. It is primarily stratigraphic, but it also indicates the lithologic character of the rocks. Some may maintain that such a system is dangerous, because it is not entirely factual. It involves interpretation and is erroneous if the field geologist is wrong in his conclusion that a slate and a biotite-garnet-sillimanite schist represent the same stratigraphic unit in different metamorphic zones. But all stratigraphic work is subject to the same kind of error, even in area of unmetamorphosed rocks. A field geologist may be in error when he correlates two belts of shale many miles apart.

Ashley *et al.* (1933, p. 431) say that lithologic constitution is the controlling basis of stratigraphic subdivisions; a formation consists of one dominant lithologic type or a repeated interlamination of two or more lithologic types. They obviously meant the present lithologic constitution and they did not anticipate the problem facing the metamorphic geologist. I would say that the basis of stratigraphic subdivision is the present lithologic constitution or the inferred lithologic constitution prior to metamorphism.

We conclude, therefore, that the study of metamorphic rocks should be primarily stratigraphic and that the present lithologic character is of secondary importance in unraveling the pre-metamorphism history.

Simultaneous Solution of Stratigraphy and Structure

It is often said that the proper way to study the geology of an area is to solve the stratigraphy first and then to decipher the structure. Nothing is further from the truth. Actually the stratigraphy and structure must be solved simultaneously. If the structure is incorrect the stratigraphy is likewise wrong. Being a structural geologist I could argue that the structure must be solved first and I might cite what at first appears to be a far-fetched example. Where strata are essentially horizontal, as in the central United States between the Appalachian Valley and Ridge Province and the Rocky Mountains, stratigraphers and paleontologists have deduced the sequence of strata and fauna on the assumption that the beds are right side up. But they have tacitly assumed that the strata have not been overturned through 180 degrees. Of course, no one would quarrel with this assumption in that area, but this example serves to indicate that the interpretation of the structure must be correct in order to establish the correct stratigraphic sequence of the formations. Geologists who have worked in faulted areas fully appreciate the difficulties encountered in working out the correct stratigraphy. As shown by E. B. Knopf (1935), failure to recognize a large fault may lead to an incorrect solution of both the stratigraphy and structure.

In highly deformed fossiliferous areas, of course, the paleontology is a powerful tool of the geologist trying to solve the stratigraphy and structure. But even here the stratigraphy and structure finaly adopted must be consistent with each other.

The metamorphic geologist, because he has so few, if any, fossils to aid him, must be all the more skillful in solving simultaneously the stratigraphy and structure. His most powerful tools are formation contacts and key beds that can be followed for considerable distances. One clear contact or key bed is more valuable than 100 petrofabric diagrams.

Even though the metamorphic geologist has successfully deduced the stratigraphic sequence and structure of a metamorphic terrain, it may be exceedingly difficult to obtain accurate data concerning the thickness of the formations. Cleavage may obscure the bedding in many places. Formations may be tectonically thickened or thinned. The present thickness may differ greatly from the thickness before deformation. Despite the many difficulties involved, the metamorphic geologist can and should make every effort to determine the present thickness of the stratigraphic units and to discover how much the thickness of any one unit differs from place to place. The thickness prior to metamorphism may be difficult to determine quantitatively unless deformed oölites, pebbles, or fossils are present (Cloos, 1947).

Of course, the metamorphic rocks may be so extensively intruded by plutonic rocks, so broken by faults, so thoroughly granitized, and so thickened and thinned tectonically that it is impossible to deduce the stratigraphy. Nevertheless, the field geologist should realize that the sedimentary rocks were laid down in a stratigraphic sequence. It is his task to use every means at his disposal to determine that stratigraphic sequence. Although these remarks may seem trite to some, it is apparent that some investigators do not approach the problem of the metamorphic rocks with this point of view.

In summarizing this part of my remarks, I emphasize again that the student of metamorphic rocks should solve the stratigraphy and structure simultaneously, because it is impossible to isolate the two from one another as though they were separate and distinct problems.

Spatial Relations of Original Sedimentary Rocks

Primary Structures

All geologists are familiar with the original structures of sedimentary rocks, such as bedding, cross-lamination, ripple marks, mud cracks, local unconformities, and related features. Shrock (1948) has recently discussed in great detail the host of primary structures present in sedimentary rocks. Investigators of metamorphic rocks should be familiar with these primary features, not only because of their intrinsic significance, but also because some of them may be confused with secondary structures.

On the other hand, there is great danger that investigators of metamorphic rocks will mis-

interpret secondary structures as primary sedimentary features. Stark and Maxson (1939) have shown that fracture cleavage has been misidentified as cross-lamination, and Ingerson (1940) has shown that minor folds have been called ripple marks. What has often been called bedding is really a secondary structure (Knopf and Ingerson, 1938, p. 214). Many of us, in our desire to find stratification, have too readily accepted interlamination of thin bands of different composition as evidence of bedding. Such interlamination may be due to the plastic injection of argillaceous material along cleavage planes transverse to the bedding. It can also result from metamorphic differentiation whereby an originally homogeneous rock becomes banded as material moves short distances in solution and segregates into layers parallel to the cleavage (Billings, 1942, p. 219).

Unconformities

Most geologists are familiar with the various kinds of unconformities and their origin. In highly deformed and metamorphosed rocks, however, folds and faults may be confused with unconformities. Strong differential movements along unconformities or the development of a pronounced transverse cleavage may conceal the existence of angular unconformities. But detailed mapping and careful observations at contacts will reveal unconformities, such as that beneath the Flag Rock formation in the Lead area, South Dakota (Noble and Harder, 1948, p. 950–951).

Contrasting Deposits of Transgressing and Regressing Seas

Many geologists working in metamorphic terrains are unfamiliar with the contrasting sequences of rocks deposited in a transgressing sea or in a regressing sea. Theoretically, if the sea invades a region of low relief, the first deposits are coarse clastics, such as gravel and sand; as the shore line moves inland, progressively finer-grained deposits are laid down at any one place, so that finer clastics, such as fine-grained sands and silt, are followed in succession by muds and finally by carbonate rocks (Lahee, 1941, p. 70–74; Sears, Hunt, and Hendricks, 1941; Stetson, 1949). Theoretically, in the deposits laid down in a regressing sea the carbonates are overlain successively by shales, fine clastics, and coarse clastics. Recognition of the sequences characteristic of transgressing and regressing seas is quite feasible in metamorphic rocks, but recrystallization may obscure those criteria based on grain size and texture. Moreover, under special conditions the sequence may be different from that normally expected.

Sedimentary Facies

In recent years a great deal of emphasis has been placed on sedimentary facies, the concept that the lithologic character of sediments deposited at one time in a single basin will differ from place to place and consequently different rocks will grade laterally into one another (Longwell, 1949). Actually this is a relatively old idea. Grabau (1921, p. 290, 399–401) enunciated it in the Appalachians, and Alpine geologists have long recognized its importance. Sloss, Krumbein, and Dapples (1949) have presented cartographic methods to express quantitatively the changes in sedimentary facies.

Like any good idea, the principle of sedimentary facies can be abused. It is a convenient device to resolve difficulties encountered in the field. Some of you are probably familiar with the old gag that a fault may be defined as the last resort of the harassed geologist. We might paraphrase this by saying that a change in facies is the first resort of the harassed stratigrapher. Faults and facies changes are both geological realities, but in some cases they are figments of the imagination. In deformed rocks what seems to be a change in sedimentary facies may actually be due to intense folding. I may cite an example from western New Hampshire (Fig. 2), where the Ammonoosuc volcanics are stratigraphically above the Albee formation, composed of quartzite and schist. In the eastern part of the township of Haverhill both the bedding and cleavage strike northeast and dip steeply (Fig. 2A). In the northern part of the area covered by the map the rocks are fine-grained biotite gneiss, amphibolite, and mica schist. Toward the southwest, quartzite and micaceous quartzite become progressively more abundant. This was at first explained by the simple device of a change in facies. Inasmuch

as the volcanics were assigned to the Ammonoosuc volcanics and the quartzites were assigned to the Albee formation, it appeared that the area illustrated a transition of one formation into another by a change in facies. Later detailed mapping on a scale of 500 feet to the inch, fortunately completed before publication, revealed numerous minor folds in the region (Billings, 1937, Fig. 5, p. 522). It eventually became clear that the contact was a single horizon that had been thrown into a series of steeply plunging anticlines and synclines (Fig. 2B). In recent years I have seen several areas where the field geologist interpreted similar transitions in metamorphic rocks as changes in sedimentary facies; they failed to make a detailed study of the structural relations at the critical contact.

It is essential that the investigator of metamorphic rocks be familiar with the manner in which sedimentary facies grade into one another. On the other hand, he must not let a superficial knowledge of this subject lead him to misinterpret field data. Moreover, his task is doubly difficult, because changes in metamorphic grade or metamorphic facies may be superimposed on changes in sedimentary facies.

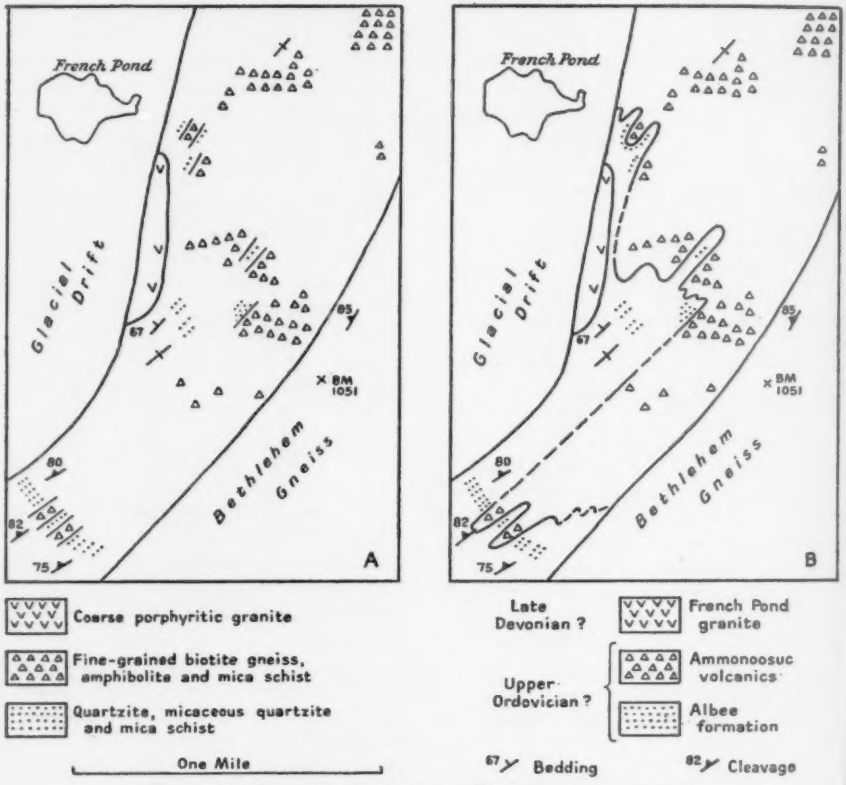

FIGURE 2.—CHANGE IN SEDIMENTARY FACIES OR INTERFOLDING?
Geologic maps of northeastern part of Haverhill, New Hampshire. A. Outcrop map, suggesting change in sedimentary facies. B. Interfingering is actually due to interfolding.

Time Surfaces

Modern stratigraphers are well acquainted with the principle that formation boundaries

may transgress what are called time lines (Lahee, 1941, p. 73). Although the principle is sound, it is difficult to apply in the field. Formation boundaries are in many cases relatively easy to locate, but by no means always. Time lines—or time surfaces would be more appropriate if we are to think in three dimensions—are more difficult to establish.

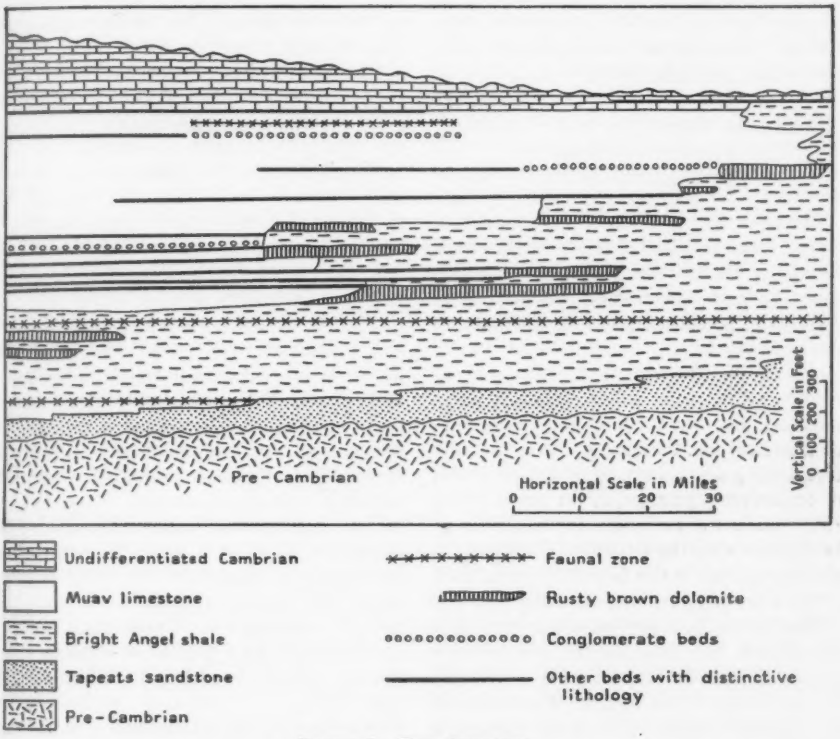

FIGURE 3.—TIME SURFACES

Associated with changes in sedimentary facies in the Grand Canyon region (after McKee, 1945; 1949).

McKee (1945, p. 24–28; 1949, p. 39) has discussed this problem, especially in reference to the strata so excellently exposed in the Grand Canyon region. As illustrated in Figure 3, he believes that the most satisfactory indicators of time surfaces are: (1) a thin zone that is characterized by distinctive fossils and that can be followed for long distances; (2) some thin distinctive bed, such as a limestone, an intraformational conglomerate, or a volcanic tuff, that can be followed for long distances; or (3) where widespread fossil zones or key beds are absent, a combination of marker units of limited extent can be used. It is assumed that the key zone or bed was deposited simultaneously everywhere. If the zone or bed transgresses the boundary between two formations it means that the contact represents different times in different parts of the area. Perhaps one of the most unequivocal types of time surface is a thin distinctive bed of volcanic ash that cannot be confused with other beds of ash. R. C. Moore (1949) cites an excellent example from the work of Bramlette and Rubey in the Upper Cretaceous rocks of Wyoming and Montana.

To establish time surfaces in metamorphic rocks is exceedingly difficult. Fossils are so rare and so deformed that we are happy if we can assign our rocks to the proper system or series. Metamorphosed ash beds are generally not

sufficiently distinctive or are so deformed that it is difficult to trace them. In eastern Vermont, however, the Standing Pond amphibolite, which can be traced for many miles and appears to be a metamorphosed volcanic tuff, transgresses the boundary between the Waits River formation and the Gile Mountain formation (Doll, 1943–1944; White and Eric, 1943).

Lithologic Character and Chemical Composition of Original Sedimentary Rocks

The inquisitive investigator of metamorphic rocks wants to know what the original sedimentary rocks were like. What was the mineralogy, texture, and chemical composition of these rocks? For many years the stratigrapher was not particularly interested in the internal constitution of the rocks with which he dealt. A shale was a shale, and a sandstone was a sandstone, and that was considered to be sufficient information. Fortunately this is no longer true, and a great deal of progress has been made in sedimentary petrography in recent years. Every student of metamorphic geology should be familiar with the information gathered by the investigators in this field (Pettijohn, 1949).

There are certain phases of stratigraphy and sedimentation, however, on which more data are needed. The mineralogical and chemical composition of the finer particles in the sedimentary rocks is still obscure. What minerals are present and how much of the material is amorphous? This query is not meant as a criticism of the men working in these fields, for I am well aware of the difficulties involved and the prodigious efforts being made to resolve them (Grim, 1942; Grim, Dietz, and Bradley, 1949). Another deficiency is the lack of adequate data on the chemical composition of sedimentary rocks. The metamorphic geologist wants to know what changes, if any, in chemical composition have occurred during metamorphism. But it is surprisingly difficult to find chemical analyses of sedimentary rocks.

As an example I may cite an experience we recently had in New England. M. T. Heald (1948), working in southwestern New Hampshire, studied in great detail some of the gneisses in the sillimanite zone of the Devonian Littleton formation. These gneisses now consist of quartz, muscovite, biotite, plagioclase, and small amounts of sillimanite, graphite, and pyrite. Both the mineralogy and chemical analyses suggest that the gneisses were pyritiferous black shales prior to metamorphism. But a search of the literature revealed few satisfactory chemical analyses of such black shales.

Paleogeography and Genesis of Original Sedimentary Rocks

The student of metamorphic rocks should go even further in his thinking. He should think of the broad paleogeographic problem. What was the nature of the basin in which the original sediments accumulated? Whence came these sediments? What was the topography and climate of the landmass whence they came? I am well aware that stratigraphers are now giving much attention to such problems (Twenhofel, 1939; Trask, 1939). Many students of metamorphic rocks do not seem to be aware that such problems exist and are completely ignorant of the criteria that may be used to obtain the answer.

These larger problems cannot be solved from studies restricted to a relatively small area. The data from many small areas must be integrated into a regional picture. Nevertheless, the geologist mapping a small area will attack his problem with much greater intelligence if he appreciates the paleogeographic problems (Krynine, 1943).

The stratigraphy and structure of a large area must make paleogeographic sense. At the moment I am not prepared to define what I mean by paleogeographic sense. I think that those stratigraphers working in unmetamorphosed sedimentary rocks will agree that even in areas of relatively simple structure the distribution of the rocks in some places is difficult to explain (King, 1949). Ultimately, however, we should have an answer to many of these puzzling problems in the distribution of sedimentary rocks.

I may cite the Paleozoic paleogeography of New England. Although a vast amount of field work remains to be done and there are several knotty stratigraphic and structural problems, enough is already known to begin to think in terms of the paleogeography. Many geologists are now working in this area, and much of their

information is unpublished. As work progresses we should constantly ask whether the proposed stratigraphy and structure fit into a reasonable paleogeography. If not, we should definitely reconsider those parts of the stratigraphy that seem incompatible with a logical paleogeography. It is obvious, however, that interpretation of the stratigraphy and structure should not be modified just to fit a preconceived paleogeography. But when sufficient data are available the stratigraphy, structure, and paleogeography should integrate into a single consistent interpretation.

Stratigraphers who work in unmetamorphosed sedimentary rocks often wonder why anyone is so foolish as to bother with metamorphic rocks. I can only remind them that their paleogeographic maps will be fragmentary and incomplete if they do not pay attention to the areas of metamorphic rocks. The Paleozoic paleogeography of the Appalachian region cannot be deduced solely from the data collected in the Valley and Ridge Province and the Appalachian Plateaus. It is equally important to understand the geology of the Piedmont, New England, and the Maritime Provinces of Canada.

Field Geology

How can the stratigraphy and structure of metamorphic areas be deduced? How can the paleogeography be unravelled? It is obvious that the answers to such problems can be found only in the field. Petrography, chemical analyses, petrofabric analysis, the investigation of the physical chemistry of silicate systems, and high-pressure studies are all helpful tools and are necessary, especially in the study of the processes of metamorphism. But solution of the structure, stratigraphy, paleogeography, and geological history requires the accumulation and analysis of a prodigious amount of field data.

This field work cannot be done rapidly. It is long and tedious. It must be done piece by piece, and integration must follow. A party of two or three geologists will take several years to study one or two hundred square miles. The local stratigraphy must be worked out in each of these areas, but to construct the broad paleogeographic setting a much larger area must be investigated. One person or a few cannot do all this work. Years of patient mapping by many people are required to gather the essential data. But all this detailed work will be in vain unless someone integrates it into a consistent and uniform whole.

I emphasize that the investigation of a large metamorphic terrain is a long and tedious task that must be done piecemeal by many people. Adequate financial support for the study of small areas is essential and publication is equally important. How can the larger philosophical problems of geology be intelligently discussed without the basic data obtained in the small areas? How shall we ever learn anything about orogeny, metamorphism, stratigraphy, paleoecology, or any other field of geology unless field investigations are encouraged?

Conclusions

It is beyond the scope of this paper to enumerate or to evaluate the kinds of field and laboratory data that the student of metamorphic rocks should seek. In a later publication I hope to consider more fully the relative values of petrofabric diagrams, measurements of lineation, measurements of the attitude of minor folds, petrography, and chemical analyses. Here it will suffice to say that, in my opinion, paraphrasing the quotation from Leigh Hunt's poem "Abou Ben Adhem:" "And lo! stratigraphy led all the rest." A more thorough appreciation of the importance of stratigraphy, sedimentation, and paleogeography is essential in the study of metamorphic rocks. My remarks are not intended to be a wholesale indictment of all those who have worked in metamorphic rocks. Some of them have done excellent stratigraphic work. But, especially during the last few decades, there has been progressively more emphasis on the petrographic, chemical, and mechanical phases of the problem. Unfortunately this has been accompanied by an increasing neglect of the stratigraphic phases of the study of metamorphic rocks.

I am sure that stratigraphers will recognize that many students of metamorphic rocks have been negligent in their duty. But stratigraphers may be cheered by the realization that much more can be done with the metamorphic rocks than has been done in the past. I also hope that

stratigraphers will recognize that in constructing paleogeographic maps they cannot ignore the metamorphic rocks. Vast areas of metamorphic rocks in the Appalachians, under the Atlantic and Gulf Coastal Plains, and in the Pacific Mountain system may be the same age as the unmetamorphosed rocks with which stratigraphers are more familiar.

It has been said that stratigraphy is the basis of all geology (Weller, 1947). This is true of the metamorphic rocks as well as the sedimentary rocks. Cognizant of the importance of petrography, physical chemistry, and structural geology, the investigator of metamorphic rocks should nevertheless give more adequate treatment to the fascinating problems in stratigraphy, sedimentation, and paleogeography that await solution in the interesting rocks to which, perhaps foolishly, he has dedicated his scientific career.

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STRATIGRAPHY AND STRUCTURE OF THE MT. PAWTUCKAWAY QUADRANGLE, SOUTHEASTERN NEW HAMPSHIRE

By Jacob Freedman

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Abstract

The chief problem in the Mt. Pawtuckaway quadrangle was to determine the relation of the early and middle Paleozoic rocks of western and central New Hampshire to supposed Pennsylvanian rocks of southeastern New Hampshire. A related problem was to determine the stratigraphic relations of the Berwick "gneiss" which Katz considered pre-Algonkian.

The Mt. Pawtuckaway quadrangle is an area of metamorphosed sedimentary rocks intruded by plutonic rocks of the late Devonian (?) New Hampshire magma series and the Mississippian (?) White Mountain magma series. The metamorphic rocks are assigned to three stratigraphic units: the Eliot, Berwick, and Littleton formations. The Eliot formation consisted originally of interbedded shale, sandstone, and dolomitic limestone; these rocks were metamorphosed to phyllite, quartz-mica schist, quartzite, and lime-silicate rocks. The Calef member at the top of the Eliot formation was originally a carbonaceous shale; it is now a dark phyllite. The Berwick formation was composed of interbedded shale, sandstone, shaly sandstone, and dolomitic limestone; these rocks have been metamorphosed to phyllite, quartzite, quartz-mica schist, and lime-silicate rock. The Littleton formation consisted originally of sandstone, shaly sandstone, and shale; these rocks were metamorphosed to quartzite, mica-schist, mica-garnet schist, mica-staurolite schist, and mica-sillimanite schist.

The New Hampshire magma series, which is rather widespread in the central and northwestern parts of the quadrangle ranges from gabbro to granite. It has been divided into five units on the basis of lithologic characteristics and distribution: Exeter diorite, quartz diorite, quartz monzonite, binary granite, and microcline granite. They are dominantly massive rocks, but locally foliation is conspicuous.

A small, circular pluton of the White Mountain magma series, ranging from gabbro to monzonite, was emplaced by forceful injection and cauldron subsidence in the Pawtuckaway Mountains in the central part of the quadrangle. Three smaller bodies of the White Mountain magma series at the southern border of the quadrangle in Raymond township were emplaced by forceful injection and volcanic eruption. They are composed of coarse-grained rocks such as gabbro, hypersthene gabbro, and hypersthene monzonite, and fine-grained rocks such as andesite, quartz latite, keratophyre, and rhyolite.

Minor folds, lineation, rare fracture cleavage and a structural terrace considered as a drag fold on Saddleback Mountain indicate that the top of the beds is dominantly to the northwest. Thus the metamorphic rocks form a northwesterly dipping homocline within the limits of the quadrangle.

Metamorphism decreases southeast and northwest from the granitic pluton. Chlorite, biotite, garnet, staurolite, and sillimanite zones are present. An actinolite isograd occurs in the biotite zone.

The Berwick formation conformably overlies the Eliot formation and underlies the Littleton formation. The Devonian Littleton formation has been traced to the Mt. Pawtuckaway quadrangle from western New Hampshire, where it is dated by fossils. Correlation with the mid-Silurian Waterville slate in Waterville, Maine, indicates that the Eliot and Berwick formations are probably Silurian.

Introduction

Location and Extent of Area

The Mt. Pawtuckaway quadrangle is located approximately 20 miles northwest of Portsmouth in southeastern New Hampshire (Fig. 1). It is bounded by parallels 43°00′ and 43°15′ north latitude, and meridians 71°00′ and 71°15′ west longitude, and is 228 square miles in area.

Problem

As geological work in New Hampshire progressed southeastward, it seemed that there must be a contact somewhere between the schists of the Littleton formation, of Lower Devonian age, and the presumably Pennsylvanian rocks of the southeastern part of the state. The Mt. Pawtuckaway quadrangle appeared a likely area in which to find such a contact and several hypotheses as to the nature of the contact were considered: (1) a conformable sequence of strata from the Devonian to the Pennsylvanian might be present; (2) the Pennsylvanian might overlie the Devonian with pronounced angular unconformity; or (3) the two groups of rocks might be in fault contact.

The stratigraphic position of the Berwick formation, which Katz (1917) considered to be a pre-Algonkian formation in the midst of Pennsylvanian

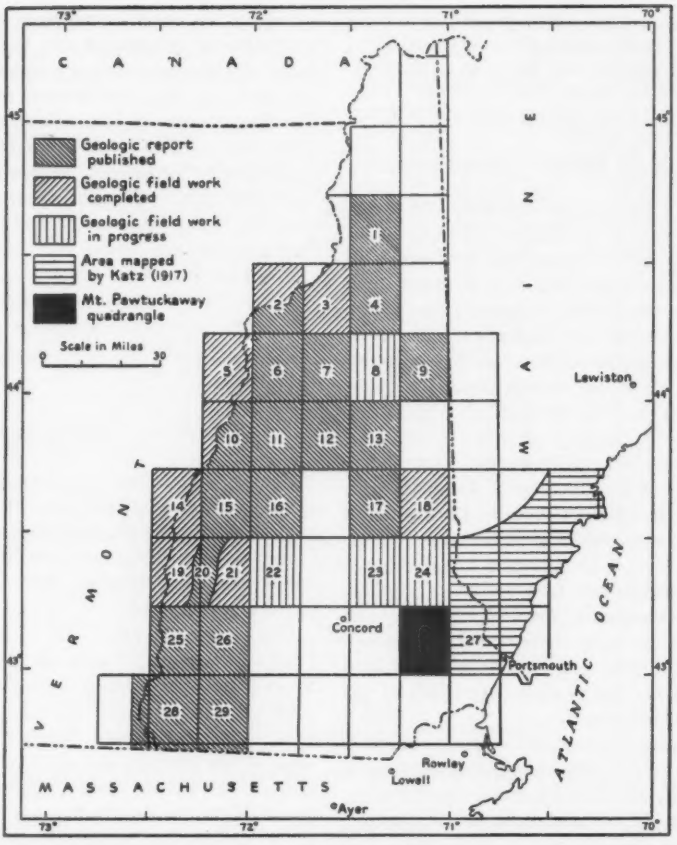


FIGURE 1.—INDEX MAP

Pawtuckaway area in solid black. Other quadrangles studied in the present investigation of New Hampshire and adjacent states are: 1. Percy (R. W. Chapman, 1948); 2. Littleton (M. P. Billings, 1937, and J. H. Eric, 1942); 3. Whitefield (R. H. Arndt, in preparation); 4. Mt. Washington (M. P. Billings, C. A. Chapman, R. W. Chapman, K. Fowler-Billings, and F. B. Loomis, Jr., 1946); 5. Woodsville (Vermont, W. S. White and M. P. Billings, in preparation); 6. Moosilauke (M. P. Billings, 1937); 7. Franconia (C. R. Williams and M. P. Billings, 1938); 8. Crawford Notch (D. M. Henderson, in preparation); 9. North Conway (M. P. Billings, 1928); 10. Mt. Cube (New Hampshire, J. B. Hadley, 1942; Vermont, J. B. Hadley, in preparation); 11. Rumney (K. Fowler-Billings and L. R. Page, 1942); 12. Plymouth (C. B. Moke, 1946); 13. Mt. Chocorua (A. P. Smith, L. Kingsley, and A. Quinn, 1939); 14. Hanover (J. B. Lyons, in preparation); 15. Mascoma (C. A. Chapman, 1939); 16. Cardigan (K. Fowler-Billings and L. R. Page, 1942); 17. Winnipesaukee (A. Quinn, 1944); 18. Wolfeboro (A. Quinn, in preparation); 19. Claremont (C. A. Chapman, in preparation); 20. Claremont-Newport area (C. A. Chapman, 1942); 21. Sunapee (C. A. Chapman, in preparation); 22. Mt. Kearsarge (C. A. Chapman); 23. Gilmanton (M. T. Heald); 24. Alton (G. W. Stewart); 25. Bellows Falls (F. C. Kruger, 1946); 26. Lovewell Mountain (M. T. Heald, in preparation); 27. Dover (T. R. Meyers, in preparation); 28. Keene and the east third of the Brattleboro, Vermont (G. E. Moore, 1949); 29. Monadnock (K. Fowler-Billings, 1949).

strata, was another problem. These stratigraphic problems made it necessary to determine the structure of the whole quadrangle, and this, in turn, led to a study of the metamorphism and the petrology.

The structural problem and the stratigraphic sequence have been solved with reasonable satisfaction. The hypotheses predicated on the Pennsylvanian age of the southeastern New Hampshire

terrane are apparently irrelevant. The strata in the Mt. Pawtuckaway quadrangle are on the northwest limb of an anticline, and hence are progressively younger to the northwest. They are older than Pennsylvanian, and are probably Silurian and Devonian.

Work in the Area and Acknowledgments

Previous work in the area has been chiefly reconnaissance. C. T. Jackson (1844) and C. H. Hitchcock (1877) studied the area briefly in their surveys of the state. F. J. Katz (1917) reported on the geology of southwestern Maine and southeastern New Hampshire (Fig. 1), but the eastern border of the Mt. Pawtuckaway quadrangle was the western limit of his published map. E. S. C. Smith (1922) made a brief study of the geology of the Pawtuckaway Mountains. The physiography and glacial geology were described by J. W. Goldthwait (1925). Silicified fault zones in Raymond and Nottingham townships were examined by T. R. Meyers (1941) in his survey of quartz deposits in New Hampshire. G. W. White (1941) examined swamps in Fremont, Epping, and Raymond townships in his search for peat deposits. C. J. Roy and the author (1944) have described the petrology and structure of the Pawtuckaway Mountains.

The present investigation is based on 9 months of field work during the summers of 1940, 1941, and 1946. Laboratory work and the preparation of the manuscript were completed during the academic year of 1946–1947.

The topographic map of the Mt. Pawtuckaway quadrangle, on a scale of 1:62,500, was enlarged to a scale of approximately 3 inches to 1 mile for mapping in the field, but the map was enlarged to 12 inches to a mile where greater detail was necessary. The laboratory work included a study of 118 thin sections of the rocks, supplemented by mineral-powder determinations under the microscope.

The author is deeply grateful to Professors Marland P. Billings and E. S. Larsen, Jr. of Harvard University for their supervision and assistance in the preparation of this paper both in the field and the laboratory. The author is indebted to Theodore Botinelly, who acted as field assistant during the summer of 1940. Professors T. R. Meyers and G. W. Stewart of the University of New Hampshire were helpful in surveying the mutual boundaries of the Mt. Pawtuckaway quadrangle and their areas of study, the Dover quadrangle to the east and the Alton quadrangle to the north respectively. The Department of Mineralogy of Harvard University supplied funds for making 97 thin sections. The Department of Geology of the University of New Hampshire loaned its collection of 21 thin sections and rocks from the Mt. Pawtuckaway quadrangle.

Funds for the colored map were provided by the New Hampshire State Planning and Development Commission and by a grant from the Penrose Bequest of The Geological Society of America.

Topography and Drainage

The Mt. Pawtuckaway quadrangle is in the seaboard lowland section of the New England physiographic province. More elevated portions of the province, belonging to the New England upland section (Fenneman, 1946), lie directly to the north and west.

Two major streams, the Isinglass and Lamprey rivers, flow eastward. The Isinglass River and its tributaries drain the northern sixth of the quadrangle. The Lamprey River, with its largest tributary, North River, drains most of the southern five-sixths of the quadrangle. The Piscassic River drains the southeast corner of the quadrangle. Swamps border most of the streams, and 36 lakes, from 0.25 mile to 2 miles across, dot the area.

The average relief is approximately 200 feet. The total relief is 1084 feet; the lowest point is 100 feet in the southeast corner of the quadrangle; the highest is 1184 feet on Saddleback Mountain in the northwest corner of the quadrangle.

The quadrangle is an area of maturely dissected hills modified by continental glaciation. The concordant summits of the hills indicate a tilted uplifted peneplain (Goldthwait, 1925, p. 5) above which the Pawtuckaway Mountains and Saddleback Mountain rise as monadnocks. Numerous drumlins throughout the area are elongated northwest-southeast. Glacial striae trend regularly S. 30°E. A line of moraines with typical knob and kettle topography may be traced through Lee, Epping, Fremont, and

Raymond townships. An outwash plain into which the Lamprey River has incised its valley extends south and southeast from the moraine.

three stratigraphic units which are progressively younger to the northwest: the Eliot and Berwick formations of probable Silurian

AGE	FORMATION	COLUMNAR SECTION	LITHOLOGY	THICKNESS IN FEET	INTRUSIVE ROCK
Quaternary			Till, glacial outwash and alluvium.		
Lower Devonian	Littleton formation		Biotite zone: gray phyllite and quartzite, with some mica schist and mica-quartz schist. Garnet zone: mica schist, mica-garnet schist, and quartzite. Staurolite zone: mica schist, mica-garnet schist, mica-garnet-staurolite schist, and quartzite. Sillimanite zone: mica schist, mica-garnet schist, mica-sillimanite schist and quartzite.	7700	White Mountain magma series (wm) Mississippian ? New Hampshire magma series (nh) middle or late Devonian
			Gove member: chiefly silvery mica-schist, some mica-staurolite schist, mica-sillimanite schist, and quartzite. 0-200 feet.		
Silurian ?	Berwick formation		Biotite zone: black, gray, and green phyllite with biotite occurring both as porphyroblasts and in groundmass; quartz-mica schist; quartzite; actinolite-quartz granulite; actinolite granulite; biotite-actinolite schist; actinolite amphibolite. Garnet zone: similar to rocks in biotite zone but some of them contain garnet..	7000	
	Eliot formation		Calef member: chiefly black phyllite; some green quartz-chlorite phyllite. 0-800 feet.	6500	
			Chlorite zone: dark-gray, light-gray, and green phyllite; quartz-sericite schist; quartzite; calcareous phyllite. Biotite zone: dark-gray, light-gray, and green phyllite with biotite porphyroblasts; quartz-mica schist with biotite porphyroblasts; quartzite; actinolite-biotite granulite.		
	Kittery formation		Thin-bedded quartzite. (not exposed in Mt. Pawtuckaway quadrangle)		

FIGURE 2.—COLUMNAR SECTION FOR THE MT. PAWTUCKAWAY QUADRANGLE

South of the village of Epping, part of the plain is covered by marine clay where the sea encroached on the land as the continental ice sheet retreated. The same sea eroded Bald Hill, a drumlin in Newmarket township, and deposited a beach and spit at its base. After recession of the ice sheet and uplift of the land, streams cut through the loosely consolidated glacial deposits and established the present poorly integrated drainage system.

General Geologic Relationships

The Mt. Pawtuckaway quadrangle is an area of metamorphosed sedimentary rocks intruded by plutonic rocks. The columnar section (Fig. 2) indicates the sequence, age, thickness and lithologic characteristics in the area with a graphical representation of the geological history. The metamorphic rocks are assigned to age, and the Littleton formation of Lower Devonian age.

Metamorphism in general decreases in intensity northwestward and southeastward from the central part of the quadrangle. The metamorphic rocks are assigned to chlorite, biotite, garnet, staurolite, and sillimanite zones. The metamorphic zones may be delineated with reasonable accuracy because the chemical composition of a sufficient, although relatively small, portion of the rock favors the precise indication of metamorphic intensity by change in mineral composition. Biotite, actinolite, garnet, staurolite, and sillimanite isograds are shown on the map of metamorphic zones (Fig. 3).

The metamorphic rocks are intruded by plutonic rocks of the Devonian New Hampshire magma series and the Mississippian (?) White Mountain magma series. Rocks of the New

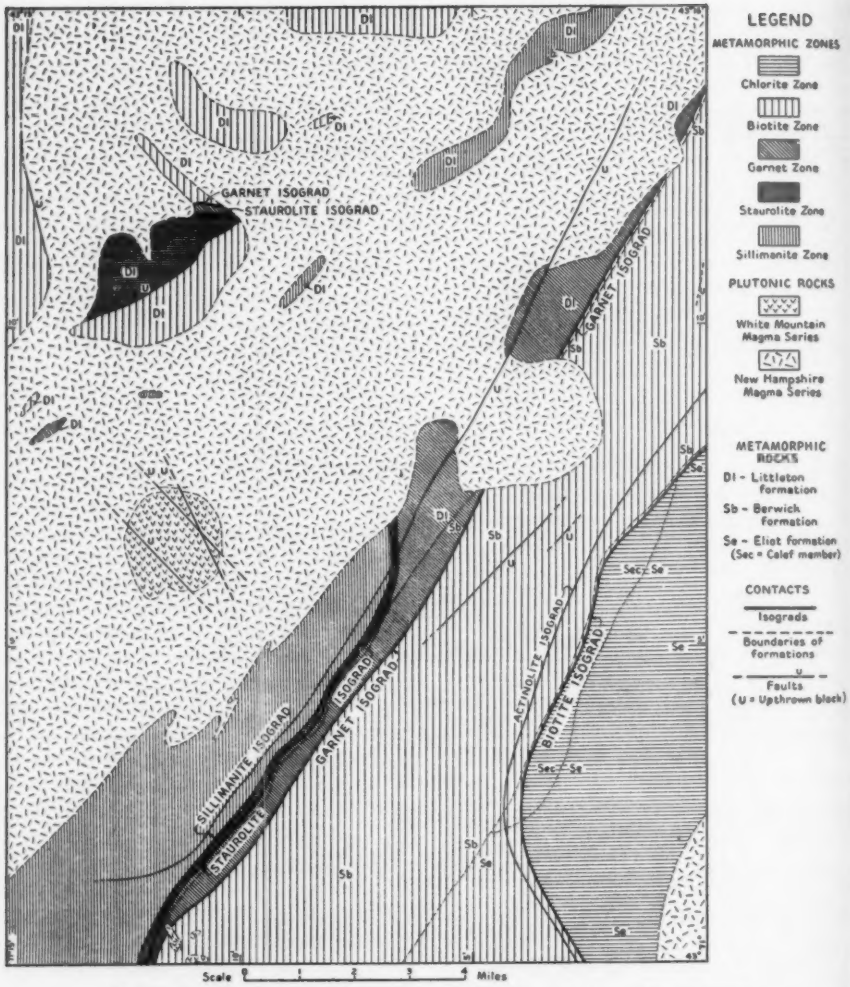


FIGURE 3.—METAMORPHIC ZONES IN THE MT. PAWTUCKAWAY QUADRANGLE

Hampshire magma series are divided into five units on the basis of lithologic characteristics, structure, and distribution. Certain rocks of the White Mountain magma series, which are lithologically divided into nine units, intrude rocks of the New Hampshire magma series as well as the metamorphic Berwick formation.

Minerals belonging to isomorphous series are referred to Winchell's (1933, Vol. II) classifications and identified through the use of his charts unless otherwise indicated.

ELIOT FORMATION (SILURIAN?)

General Statement

The Eliot formation occupies a northeasterly trending belt in the southeast part of the quadrangle (Pl. 1). It is overlain to the northwest by the Berwick formation at a gradational contact, and in the Dover quadrangle to the east, it conformably overlies the Kittery quartzite.

The Eliot formation (Katz, 1917, p. 169)

was named from the town of Eliot, Maine, which lies 10 miles east of the eastern edge of the Mt. Pawtuckaway quadrangle. Katz (1917, Plate 61) shows two belts of the Eliot formation, the more westerly of which is continuous with the belt mapped in the Mt. Pawtuckaway quadrangle. Although Katz called the formation the Eliot slate, it seems better to revise the name to Eliot formation inasmuch as true slates are nowhere abundant. The age of the formation is considered on a later page.

Lithologic Characteristics

Most of the Eliot formation in the Mt. Pawtuckaway quadrangle lies in the chlorite metamorphic zone and consists chiefly of dark-gray phyllite, light-gray phyllite, green phyllite, quartzite, and quartz-sericite schist. Although calcareous phyllites are not exposed in the Mt. Pawtuckaway quadrangle, they are undoubtedly an integral part of the formation because they are not uncommon in the Dover quadrangle to the east. Their failure to crop out in the Mt Pawtuckaway quadrangle is probably due to poor exposures.

In the southern part of the quadrangle, in Brentwood township and the northeast corner of Fremont township, the Eliot formation lies southwest of the biotite and actinolite isograds (Fig. 3); here the phyllites contain biotite porphyroblasts and the calcareous phyllites have been metamorphosed to actinolite granulite.

The phyllites (Table 1, Cols. 1, 2) constitute 65 per cent of the main part of the Eliot formation in this area. Some are dark-gray to black, fine-grained, crinkled rocks which weather black or rusty brown. Others are gray and still others, owing to the abundance of chlorite, are green. In general, bedding is better exposed where more quartzose beds are interbedded at intervals of a few inches to a few feet. The principal minerals are sericite, quartz, and chlorite. There are minor amounts of magnetite, apatite, and carbonaceous material. The phyllites have a well-developed foliation caused by the parallel arrangement of the plates of sericite and chlorite. The chlorite is characterized by abnormal blue interference colors and is probably penninite.

The phyllites southwest of the biotite isograd in Brentwood and Fremont townships are similar except for the presence of biotite porphyroblasts parallel to the schistosity averaging 2 mm. in diameter.

The quartz-mica schists constitute about 10 per cent of the formation. They are light-gray to dark-gray, brown weathering, fine-grained to coarse-grained rocks. In the chlorite zone they are quartz-sericite schists composed essentially of quartz, sericite, and chlorite, with minor amount of sphene. A well-developed foliation is produced by the parallel arrangement of platy minerals. Southwest of the biotite isograd in Fremont and Brentwood, biotite occurs both as porphyroblasts and as an integral part of the groundmass (Table 1, column 3).

Table 1. Approximate Modes*—Eliot Formation

	1	2	3
Number of thin sections	2	1	1
Quartz	22	75	65
Sericite	68	5	
Muscovite			2
Biotite			30
Chlorite	8	20	2
Sphene			1
Magnetite	tr		tr
Apatite			tr
Carbonaceous matter	2		
Range of grain size in mm.	0.01–1.00	0.1–1.0	0.5–1.0

1. Sericite-quartz phyllite.
2. Quartz-chlorite phyllite.
3. Quartz-mica schist.

* Percentages indicate volume per cent.

The quartzite is a light-gray to greenish, massive, fine-grained rock constituting approximately 15 per cent of the formation. Weathered exposures indicate that these rocks are thinly bedded in some localities. No thin sections of quartzite were studied; they are megascopically similar to quartzites in the Berwick formation which are described below.

The actinolite granulite, confined to that part of the formation in Brentwood and Fremont lying southwest of the actinolite isograd (Fig. 3), constitutes about 10 per cent of the outcrops. It is a bedded, light-gray to dark-gray, fine-grained to medium-grained rock with porphyroblasts of light-green actinolite and biotite, quite like similar rocks in the Berwick formation.

The Calef member at the top of the Eliot formation, which forms a narrow belt between the villages of Lee and Epping, is composed almost exclusively of black phyllite. It is named from the Calef Road which now utilizes the abandoned right-of-way of the Worcester, Nashua, and Portland division of the Boston and Maine Railroad shown on the accompanying map (Pl. 1).

Thickness

The true thickness of the Eliot formation is impossible to determine in the Mt. Pawtuckaway quadrangle because the lower part of the formation is cut out by the Exeter diorite. The Eliot formation is 6,500 feet thick. The Calef member attains a maximum thickness of 800 feet but it is locally absent.

Berwick Formation (Silurian?)

General Statement

The Berwick formation trends northeast from the southern boundary of the quadrangle in Fremont and Brentwood townships to the eastern boundary in Barrington township. The Berwick formation lies stratigraphically above the Eliot formation and underlies the Littleton formation. Contacts between these formations are gradational. The formation was named by Katz from the town of Berwick, Maine, which lies 7 miles northeast of the northeastern corner of the Mt. Pawtuckaway quadrangle. The belt in the Mt. Pawtuckaway quadrangle can be traced continuously to the type locality (Katz, 1917, Pl. 61). Katz (1917, p. 166) called the formation the Berwick gneiss but as the rocks show good bedding, rather than a gneissic structure, the name Berwick formation is suggested and will be used in this paper.

Lithologic Characteristics

The Berwick formation in the Mt. Pawtuckaway quadrangle lies chiefly in the biotite metamorphic zone and consists chiefly of black, gray, and green phyllites with biotite occurring both as porphyroblasts and in the groundmass; it also includes quartz-mica schist, quartzite, actinolite-quartz granulite, actinolite granulite, biotite actinolite schist, and actinolite amphibolite. The extreme western part of the formation is in the garnet zone (Fig. 3). The best exposures of the Berwick formation are found along Calef Road between the Strafford-Rockingham county line and Wheelwright Pond, along the South Side State Road between Raymond and Bunker Pond, and in the vicinity of Jones Hill.

The phyllites, constituting about 35 per cent of the formation, are black to dark-gray, rusty brown-weathering rocks composed essentially of biotite, chlorite, sericite (or muscovite) and quartz. Some of the biotite and chlorite occurs as porphyroblasts.

One variety, comprising about 10 per cent of the Berwick formation, may be termed sericite-biotite-quartz phyllite (Table 2, column 1) and consists of dark-gray, crinkled, fine-grained, schistose rocks. Most common near the base of the Berwick formation southwest of Lee Station, they are also found interbedded with sericite-chlorite-biotite phyllite at Wheelwright Pond. The sericite-biotite-quartz phyllites occur in beds which range from a fraction of an inch to a few inches thick and are generally interbedded with quartzites. In thin section, they show intense minor folding and consist essentially of sericite, biotite, chlorite, muscovite, and quartz in flakes and grains with an average diameter of 0.05 mm. Carbonaceous material is disseminated in small patches and there are veinlets and grains of calcite.

A second variety, which may be called sericite-chlorite-biotite phyllite (Table 2, column 2), comprises about 5 per cent of the formation. Chlorite porphyroblasts, averaging 1 mm. in diameter, are set in a fine-grained, black to dark-gray schistose groundmass. The beds are a few feet thick. These rocks crop out on Calef Road at Wheelwright Pond in Lee. Microscopic study shows that the chlorite porphyroblasts have the abnormal blue interference color characteristic of penninite. The groundmass consists of sericite, biotite, and quartz, with small quantities of calcite and carbonaceous material.

A third variety, which may be termed biotite phyllite (Table 2, column 3), constitutes 20 per cent of the formation. Biotite porphyroblasts 1 mm. in diameter are set across the foliation in a light-gray to dark gray schistose groundmass. These rocks crop out chiefly in the vicinity of Calef Road north of Lee Station and are interbedded with quartzites, sericite-chlorite-biotite phyllites, and sericite-biotite-quartz phyllites. The groundmass is schistose and consists primarily of quartz and sericite in grains and flakes about 0.01 mm. in diameter.

Table 2. Approximate Modes—Berwick Formation

TABLE 2. APPROXIMATE MODES—BERWICK FORMATION

	1	2	3	4	5	6	7	8	9
Number of thin sections	4	3	5	3	3	2	4	4	5
Quartz	55	8	38	70	83	71	44	49	
Actinolite						15	29	15	60
Biotite	11	13	51	23	7	10	1	15	8
Clinozoisite							10		
Chlorite	7	15			5		tr		6
Muscovite	7			7	2				9
Sericite	19	62	8		tr				
Orthoclase	2								10
Microcline						2		7	
Plagioclase			2			1	14	13	1
Magnetite	tr	tr	tr	tr	tr	tr			tr
Ilmenite									tr
Apatite	tr		tr	tr	tr	tr	tr	tr	tr
Sphene		tr	tr		tr	tr	1	1	1
Limonite					tr				tr
Tourmaline			tr						5
Staurolite									tr
Garnet							1		
Calcite	tr	1			3	tr	tr	tr	
Zircon	tr	tr		tr	tr		tr	tr	
Pyrite	tr	tr	tr	tr			tr	tr	
Carbonaceous matter		1							
Range in anorthite content in plagioclase			28				27–45	24–65	
Range of grain size in mm	0.02–0.5	0.02–0.5	0.01–1.0	0.1–1.0	0.05–0.2	0.2–0.5	0.03–5.0	0.2–0.35	1.0–25.0

1. Sericite-biotite-quartz phyllite.
2. Sericite-chlorite-biotite phyllite.
3. Biotite phyllite.
4. Quartz-mica schist.
5. Quartzite.
6. Actinolite-quartz granulite.
7. Actinolite granulite.
8. Biotite-actinolite schist.
9. Actinolite amphibolite.

Quartz-mica schist (Table 2, column 4), comprising about 20 per cent of the Berwick formation, is a dark-gray to black, fine-grained to coarse-grained rock. Some beds are more quartzose than others and consequently most exposures show very good bedding. The quartz-mica schist grades with increase of sericite toward the phyllites and with increase of quartz toward the quartzites. The quartz-mica schist crops out in the central and northwestern parts of the Berwick formation.

A microscopic study indicates that quartz ranges from 65 to 75 per cent of the rock. The quartz is generally granoblastic, but in some thin sections it is slightly cataclastic. Large grains are broken down to tiny irregular grains and the larger grains show undulatory extinction. Some grains of quartz and feldspar are poikiloblastic in biotite. Biotite occurs in irregular grains as porphyroblasts parallel to the foliation and in continuous foliation planes. It ranges from 20 to 25 per cent of the rock. Pleochroic haloes around euhedral and subhedral zircon in biotite are numerous. Unaltered chlorite is present in some flakes of biotite. The γ index of the biotite ranges from 1.628 to 1.648. A biotite with these indices has approximately 45 to 65 per cent of the annite-siderophyllite molecule. Muscovite and sericite make up as much as 10 per cent of the rocks and occur as fine-grained flakes interlayered with biotite and quartz, and as coarse-grained flakes with subparallel orientation, cutting slightly across and replacing biotite. The coarse-grained muscovite, $2V = 38°$ and $\beta = 1.592$, is essentially a mica with 40 per cent each of the magnesian and potash muscovite molecules and 20 per cent of the ferric iron muscovite molecule. Accessory minerals are magnetite, hematite, apatite, pyrite, zircon, tourmaline, sphene, anatase, and carbonaceous matter.

The quartzites (Table 2, column 5), which comprise about 20 per cent of the formation, are greenish, bluish-gray, and black, fine-grained, massive to thinly-bedded rocks. The beds range in thickness from a fraction of an inch to a few feet. Quartzite laminae as thin as 0.04 inch, with micaceous partings, are found in some rocks, though more commonly the quartzite laminae are 0.1 inch thick. The best exposures of quartzite occur along Calef Road between Lee Station and Wheelwright Pond, and in the bed of North River at its junction with the road west of Lee Station.

Any rock containing 80 or more per cent quartz has been arbitrarily designated a quartzite. The amount of quartz estimated in 3 thin sections ranges from 80 to 85 per cent. The quartz is generally granoblastic, but a number of specimens have cataclastic texture. Most of the grains are dotted with tiny inclusions, and some have undulatory extinction. Biotite, sericite, and chlorite occur as irregular grains averaging 0.05 mm. in diameter. They range from a random orientation to perfect parallelism. Zircon grains surrounded by pleochroic haloes are embedded in the biotite. The chlorite, which gives a greenish tinge to the rocks, is ripidolite with end member proportions of 33 per cent antigorite and 67 per cent amesite. Accessory minerals are anhedral to subhedral magnetite, apatite and sphene. Scattered and interstitial grains of calcite make up as much as 10 per cent of the rock

Actinolite-quartz granulite (Table 2, column 6) composes approximately 5 per cent of the Berwick formation. Good exposures of actinolite-quartz granulite crop out on the banks of the Lamprey River at Epping village and northwest of North Epping. They are fine-grained, gray or greenish-gray, granular, massive to faintly bedded rocks. Beds range from a few inches to a few feet thick. Actinolite occurs as tiny needles about 1.0 mm. long scattered through the rock or locally in thin layers composed of numerous, diversely oriented needles.

Outcrops of actinolite granulite, which constitute about 5 per cent of the formation, occur chiefly in Fremont on the road west of Lyford Crossing, and on the South Side State Road 1 mile east of Raymond. Actinolite granulite (Table 2, column 7) occurs in greenish-white to gray beds and in lenses which range from 1 to 3 inches thick. The rocks are granular with dark-green to black actinolite grains scattered throughout the rock or concentrated in layers up to 0.1 inch thick. The average grain size of the groundmass is about 0.03 mm., whereas porphyroblasts of actinolite and garnet are about 5.0 mm. in diameter.

Biotite-actinolite schist, composing 10 per cent of the formation, is interbedded with thin layers of actinolite granulite. In some outcrops, the actinolite granulite occurs in the center of the lime-silicate bed, flanked by biotite-actinolite schist, which grades into normal quartz-biotite schist of the Berwick formation. The beds of biotite-actinolite schist have a minimum thickness of only 0.2 inch, but most beds range up to 1 foot thick. The biotite-actinolite schist (Table 2, column 8) is a dark-gray to black, medium-grained to coarse-grained, foliated rock with biotite and actinolite aligned in the foliation planes. The average grain size is 0.5 to 3.0 mm.

Actinolite amphibolite (Table 2, column 9), composing less than 5 per cent of the formation, is a distinctive rock that crops out at four places. Two outcrops occur on Calef Road: one opposite the southwestern end of Wheelwright Pond and the other at the second road northeast of South Lee.

A third outcrop of actinolite amphibolite occurs on the road north of Kennard Hill in Epping, 1 mile west of the road junction. The fourth is on the road at the foot of the hill 0.5 mile east of Flint Hill in Raymond. It is a coarse-grained, light-greenish-gray, massive to foliated rock composed dominantly of bundles of fibrous actinolite. The maximum grain size of the actinolite bundles is 25.0 mm., but the average is 1.0 to 3.0 mm. The beds range from 6 inches to 3 feet thick.

Mineralogy of the Lime-silicate Rocks

The amphiboles, comprising from 5 to 80 per cent of the rocks by volume are actinolite and actinolitic hornblende. They are pleochroic with X = colorless to pale yellow, Y = pale greenish-yellow, Z = pale green. The indices for actinolite range as follows: $\alpha = 1.618–1.638$, $\beta = 1.631–1.655$, $\gamma = 1.642–1.660$, and the extinction angle $Z \wedge c$ ranges from 11° to 20°. Actinolites with these indices have between 12 and 38 per cent of the ferrotremolite molecule $CaFe_5Si_8O_{22}$, by weight. The actinolitic hornblendes have higher indices and maximum extinction angles: $\alpha = 1.645–1.650$, $\beta = 1.659–1.666$, $\gamma = 1.669–1.672$, $Z \wedge c = 15°–20°$, and $2V = 80°$, negative. These data indicate that the hornblende contains approximately 40 per cent by weight of the $CaFe_5Si_8O_{22}$ molecule. The actinolite and actinolitic hornblende are similar to those in lime-silicate rocks elsewhere in New Hampshire. The amphibole occurs as laths, irregular patches, and, in the actinolite amphibolite, as sheaf-like aggregates. A few grains are euhedral, but most are anhedral. Quartz and feldspar are poikiloblastic in and embay the amphibole. Pleochroic haloes around zircon grains embedded in the amphibole are not uncommon.

Diopside was found in only one of the lime-silicate rocks.

As much as 20 per cent biotite occurs in the lime-silicate rocks. The pleochroic formula for this biotite is: X = pale yellow, Y and Z = brown. The γ index ranges from 1.628–1.648. This indicates that the biotite is intermediate between the magnesian and iron ends of the series, and that the annite-siderophyllite molecules constitute 45 to 65 per cent of the mineral. This biotite is richer in iron than biotite in lime-silicate rocks found elsewhere in New Hampshire. Biotite occurs as laths, flakes, and in irregular patches. Remnants of unaltered chlorite suggest incomplete equilibrium in some rocks. Quartz and feldspar grains occur poikiloblastically in biotite, and pelochroic haloes surround grains of zircon in biotite.

In only two of the lime-silicate rocks is chlorite found in quantities greater than a trace. In these, the chlorite, making up 20 per cent of the rock, is a clouded green, rusty alteration of actinolite, with approximately 2 per cent of green, faintly pleochroic, radially fibrous clinochlore in serpentine shapes.

The garnets from the lime-silicate rocks have an index of refraction of 1.778; a reliable specific gravity could not be obtained so an X-ray powder picture was made to determine the unit cell. The combined information indicates that the garnet is composed of 70 per cent pyrope + almandite, and 30 per cent grossularite + andradite.

Although orthoclase and microcline range from a trace to 35 per cent of the rocks in a few specimens, plagioclase is the dominant feldspar in the lime-silicate rocks. The potash feldspars are slightly sericitized or kaolinized in most specimens. The plagioclase feldspars range from oligoclase, An_{24}, to labradorite, An_{65}, and from a few scattered grains to 30 per cent of the rock. A few grains of albite variety cleavelandite, were recognized in one specimen by their curved cleavage. Both potash and plagioclase feldspar grains occur with either quartz or actinolite grains in masses, layers, and lenses. The feldspars are also poikiloblastic in actinolite.

The most abundant accessory minerals are ilmenite, magnetite, apatite, sphene, zircon, and pyrite; limonite, leucoxene, and calcite are not uncommon.

Thickness

Katz, in considering the age of the Berwick "gneiss" as pre-Algonkian, thought that the bottom of the formation was not exposed. The thickness of the formation would therefore not be determinable. The top and bottom of the formation are shown on the Mt. Pawtuckaway map (Pl. 1). A maximum estimate of the thickness of the Berwick formation is 7,000 feet.

Littleton Formation (Lower Devonian)

General Statement

The metamorphic rocks in the northwestern two-thirds of the Mt. Pawtuckaway quadrangle are assigned to the Lower Devonian Littleton formation. The schists belonging to the Littleton formation have been traced from the type locality (Billings, 1937) southeast across New Hampshire to the southern edge

of the Winnipesaukee and Wolfeboro quadrangles (Billings, 1945, p. 42). G. W. Stewart (oral communication) has informed the present writer that the schists of the Littleton formation can be traced southward through the Alton quadrangle (Fig. 1) to the northern boundary of the Mt. Pawtuckaway quadrangle. Katz (1917) assigned to the Gonic formation schists, in the Dover quadrangle to the east, that are continuous with the Littleton formation in the Mt. Pawtuckaway quadrangle.

The Littleton formation in the Mt. Pawtuckaway quadrangle is disrupted by plutonic rocks so that many of the exposures occur as islands in a sea of granitic rocks. However, northeast of the Mt. Pawtuckaway quadrangle, Katz (1917) demonstrated that these strata constitute a single continuous unit.

Lithologic Characteristics

General Statement.—Although the rocks of the Littleton formation in the Mt. Pawtuckaway quadrangle were initially rather limited in composition, consisting chiefly of arenaceous shale and sandstone, the rocks now show a considerable range in mineral composition because of differences in the grade of metamorphism. The belt extending northeast from the southwest corner of the quadrangle now lies in the garnet, staurolite, and sillimanite zones (Fig. 3). Three areas now isolated in the midst of plutonic rocks are in the sillimanite zone: northeast of West Barrington, a mile south of Northwood, and 1 mile east-northeast of Deerfield. The area northwest of the fault on Saddleback Mountain is in the staurolite zone. The area completely surrounded by plutonic rocks 1 mile southwest of Deerfield is in the garnet zone. Five other areas are in the biotite zone: northeast of Strafford Post Office, Oak Hill, southeast of the fault on Saddleback Mountain, 1 mile west of Deerfield Parade, and the long belt extending 6 miles along the northern part of the western edge of the quadrangle. This great range in the grade of metamorphism necessitates rather extensive descriptions of the rocks constituting the formation.

The name Gove member is introduced here for a distinctive, narrow belt of mica-staurolite schist and mica-sillimanite schist exposed between Nottingham Square and Raymond. The name is taken from Gove School, 2 miles northeast of Raymond.

Biotite zone.—In the biotite zone in the northwestern part of the quadrangle, the Littleton formation consists chiefly of phyllites and quartzites with minor amounts of mica schist and mica-quartz schist. Bedding is conspicuous due to the alternation of rock types which resist erosion with variable degrees of success. The beds range from a fraction of an inch to a few inches thick, but the average bed is 1–2 inches thick.

The phyllites constitute 25 per cent of the formation in the biotite zone. They are silvery gray, schistose, fine-grained rocks with crinkles or small open folds similar to those found in the Calef member of the Eliot formation. Schistosity is characteristically parallel to the bedding, but locally a faint fracture cleavage is inclined to the bedding. The chief minerals are quartz and sericite. A mode of a quartz-rich variety is given in Table 3, column 1.

Mica schists form 25 per cent of the formation in the biotite zone. They are gray, medium-grained, schistose rocks and differ from the phyllites in that they have a coarser grain. The essential minerals are muscovite, quartz, biotite, and chlorite (Table 3, column 2).

The quartz-mica schists and quartzites constitute 50 per cent of the formation in the biotite zone. They are gray rocks containing more than 80 per cent quartz. Whereas the quartz-mica schists are schistose, the quartzites are relatively massive. The essential minerals are quartz, biotite, and muscovite (Table 3, column 3).

Garnet zone.—The principal rocks are mica schist, mica-garnet schist, and quartzites.

Mica schists (Table 3, column 4), 55 per cent of the formation in the garnet zone, are medium-grained to coarse-grained, dark-gray, thin-bedded, schistose rocks. The average thickness of the beds is an inch, and the average grain size is 1.0 mm. with a maximum of 10.0 mm. and a minimum of 0.01 mm. This rock differs from the quartzite in having a smaller percentage of quartz, a greater percentage of mica and a definite schistose structure.

The mica schist is seen under the microscope to consist essentially of quartz, biotite and plagioclase, with small percentages of muscovite, orthoclase, microperthite, and chlorite. Accessory minerals are sphene, zircon, apatite, magnetite, tourma-

TABLE 3. APPROXIMATE MODES—LITTLETON FORMATION

TABLE 3. APPROXIMATE MODES—LITTLETON FORMATION

	1	2	3	4	5	6	7	8	9	10
Number of thin sections	1	3	2	6	3	3	5	7	1	4
Quartz	75	35	85	45	34	80	41	57		33
Sericite	23	tr			2					
Muscovite		50	5	1	34	2	20	10		
Biotite		10	10	32	18	7	8	10		5
Chlorite		10	tr	tr		2		7		
Magnetite		tr	tr	tr	tr	tr	1	tr		
Apatite		tr		tr	tr	tr	tr	tr		tr
Tourmaline	2	tr	tr	tr	tr	1	1	tr		
Garnet			tr		10		2	1		
Zircon			tr	tr	tr		tr	tr		tr
Orthoclase		5		tr						
Limonite		tr								
Plagioclase				22	2	7		1		14
Staurolite					tr		27	tr		
Sillimanite								13		
Sphene				tr		tr				1
Ilmenite					tr		tr	tr		
Pyrite								tr		tr
Rutile								tr		
Amphibole						1				
Actinolite										47
Diopside									40	
Scapolite									60	
Range in anorthite content in plagioclase				30–68		27–50				27–62
Range of grain size in mm	0.1–0.5	0.5–2.0	0.2–1.0	0.01–10.0	0.5–2.0	0.1–5.0	0.5–50.0	0.5–10.0	0.5–2.0	0.1–2.0

1. Phyllite.
2. Mica schist.
3. Micaceous quartzite.
4. Mica schist.
5. Mica-garnet schist.
6. Quartzite.
7. Mica-staurolite schist.
8. Mica-sillimanite schist.
9. Diopside-scapolite granulite.
10. Actinolite granulite.

line, and limonite. Quartz grains are equigranular to inequigranular, and granoblastic with mosaic or sutured edges. Undulatory extinction ranges from slight to strong. Numerous tiny dark inclusions are scattered irregularly through the mineral. Quartz embays biotite and feldspar. The average grain size of the quartz is 0.1 mm. and it ranges from 5 to 68 per cent in volume in the mica schists. Biotite occurs in stubby oriented flakes separating layers of quartz and feldspar. The maximum grain size of the biotite is 2.0 mm. Small euhedral to subhedral grains of zircon, surrounded by pleochroic haloes, and apatite are included in biotite. Biotite in quartz-mica schists ranges from 20 to 50 per cent by volume.

Plagioclase is the dominant feldspar and occurs in most rocks; orthoclase is a minor mineral in a few. The volume of plagioclase in mica schist ranges from 1 to 40 per cent. Composition ranges from albite, An_8, to labradorite, An_{68}. Numerous subhedral quartz grains embay and are poikiloblastic in plagioclase. Most of the plagioclase is fresh but in some rocks it is sericitized. Muscovite is a late mineral, partially replacing chlorite and biotite and forming prophyroblasts perpendicular to the foliation. Chlorite occurs in only two rocks, both of which may have been affected by mineralizing fluids related to a silicified fault zone. It is essentially diabantite, a chlorite containing approximately 60 per cent of the amesite molecule.

Mica-garnet schist (Table 3, column 5), about 20 per cent of the formation in the garnet zone, is a light-gray to dark-gray, coarse-grained, schistose rock with scattered prophyroblasts of garnet. It is similar to the mica schist but has garnet, less quartz, more mica, and is coarse-grained. Maximum grain-size is 1.0 mm. for the groundmass and 2.0 mm. for the porphyroblasts. The beds range from a few inches to 10 to 15 feet thick. The best exposures crop out on the shore of Jones Pond (now known as Onway Lake) in the west-central part of Raymond township.

The following essential minerals have been observed under the microscope: muscovite, biotite, quartz, and garnet. Some plagioclase feldspar is also present. Accessory minerals are tourmaline, zircon, apatite, ilmenite, and magnetite. Quartz is granoblastic to cataclastic. Most grains show strong undulatory extinction. Muscovite occurs in long, narrow, wavy flakes replacing biotite. According to the optical data—$\beta = 1.583$, $2V = 36°$—it is muscovite with 64 per cent of the potash molecule, 34 per cent of the magnesian, and 2 per cent of the ferric iron molecule. Biotite is scattered with irregular wispy flakes. Chlorite, because it is associated with higher intensity minerals, is apparently of retrograde origin. Zircon grains, surrounded by pleochroic haloes, and anhedral magnetite are poikiloblastic in biotite. Almandite garnet grains are euhedral to anhedral. Tourmaline and quartz grains occur poikiloblastically and chlorite, muscovite and magnetite fill fractures across the garnet. Plagioclase is not abundant in the rocks. In one specimen, albite, variety cleavelandite, is the plagioclase.

The quartzites (Table 3, column 6) making up 25 per cent of the formation in the garnet zone, are light-gray, massive to thin-bedded, faintly foliated, fine-grained to coarse-grained rocks with a vitreous luster. Beds range from a fraction of an inch to a few inches thick.

Microscopically, quartz is the essential mineral with plagioclase, biotite, garnet, staurolite, amphibole, pyroxene, muscovite, and chlorite present in quantities up to 8 per cent. The accessory minerals are zircon, apatite, calcite, sphene, and tourmaline. The amount of quartz ranges from 80 to 90 per cent of the rock and the grain size of the quartz ranges from 0.01 mm. to 1.0 mm. Most of the quartz is equigranular, granoblastic, but grains with sutured edges have been observed in some specimens. Undulatory extinction and tiny inclusions in the quartz are common.

Plagioclase makes up as much as 8 per cent of most of the quartzites. It ranges from oligoclase, An_{25}, to andesine-labradorite, An_{50}. Carlsbad and albite twinning characterize the mineral in most thin sections. Biotite is scattered and ranges from 2 to 5 per cent of the quartzites, and from 0.5 mm. to 2.0 mm. in diameter. Generally it occurs in stubby, oriented flakes with zircon inclusions surrounded by pleochroic haloes. A little residual chlorite remains unaltered in some of the biotite flakes. Garnet and staurolite occur as porphyroblasts in a few specimens with an average grain size of 2.0 mm. for the former and 5.0 mm. for the latter. The amphibole and pyroxene occur in a few rocks as small scattered, euhedral to subhedral grains. Muscovite flakes cut across grains of quartz and biotite, indicating that the muscovite has formed later.

Staurolite zone.—The principal rocks are mica schists, mica-garnet schists, and quartzites similar to the rocks found in the garnet zone. The presence of mica-staurolite schist distinguishes the staurolite zone from the garnet zone.

Mica-staurolite schist (Table 3, column 7) constitutes approximately 20 per cent of the Littleton formation in the staurolite zone. There are two distinct types. The first crops out in the narrow belt of the Gove member of the Littleton formation

between Raymond and Nottingham Square. The groundmass is silvery, crinkly to broadly-folded muscovite, and the porphyroblasts are garnet and staurolite. The euhedral to subhedral staurolite porphyroblasts range up to 10 mm. long. Thin layers of mica-staurolite schist are interbedded with mica schist. The second type, typical of Saddleback Mountain, has less muscovite and more quartz and occurs in foliated rocks with biotite and staurolite porphyroblasts. Staurolite is more abundant and occurs as single and twin crystals scattered and in aggregates along planes. The twin plane is (232), with individuals crossing at 60°, forming "X" shaped twins. Bedding is obscure, but the strata range from a fraction of an inch to several feet in thickness. The average grain size of the staurolite porphyroblasts is 2.0 mm., but the maximum is as much as 50.0 mm.

In thin section, the essential minerals are staurolite, muscovite and quartz. Biotite and garnet are present in some specimens. Accessory minerals are tourmaline, magnetite, apatite, zircon, and ilmenite. Staurolite occurs as euhedral to subhedral porphyroblasts in which rounded quartz grains are poikiloblastic. Cracks filled with quartz, sericite, and biotite cut the porphyroblasts; anhedral grains of magnetite are scattered through the rock. The optical properties of staurolite are: positive, $2V = 88°$, $\alpha = 1.739$, $\beta = 1.745$, $\gamma = 1.751$. The optical properties and related chemical properties are remarkably uniform. Muscovite occurs in thin foliation planes separating lenses of quartz. In some specimens, it replaces biotite with the formation along cleavage traces of ilmenite partly altered to leucoxene. Quartz grains are generally granoblastic and in part cataclastic. Undulatory extinction may be observed in the quartz grains in most specimens. Biotite occurs in irregular flakes replaced by muscovite in part or completely with the formation of magnetite and ilmenite along the cleavage traces. Zircon grains surrounded by pleochroic haloes are included in biotite. A few euhedral crystals of garnet with an average grain size of 2.0 mm. enclose quartz and biotite in numerous fractures and poikiloblastically.

Sillimanite zone.—The rocks of the Littleton formation are characteristically coarser-grained in the sillimanite zone than in the other zones. The principal rocks, mica schist, mica-garnet schist, and quartzite, are mineralogically identical with similar rocks found in the garnet zone. This zone is distinguished from the garnet zone by the presence of mica-sillimanite schist. Diopside-scapolite granulite and actinolite granulite are local types.

Microscopically, the sillimanite-bearing rocks are similar to those mentioned above. The essential minerals are sillimanite, quartz, biotite and muscovite. Chlorite, staurolite, garnet and feldspar are also found in some specimens. The accessory minerals are ilmenite, magnetite, tourmaline, zircon, and pyrite. Sillimanite ranges in the schistose rocks from a trace to 15 per cent and makes up 55 per cent of the sillimanite schist of Mt. Misery in the west-central part of Barrington township. It occurs chiefly as long slender needles or fibrous sheaves of needles in quartz. Sillimanite also forms needles on the ragged edges of biotite flakes. In the sillimanite schist it occurs as thick bundles of fibers and needles which bend and swirl around quartz and garnet grains. The indices of the sillimanite are lower than those for most sillimanite. In these rocks $\alpha = 1.645$ and $\gamma = 1.660$, whereas elsewhere $\alpha = 1.638$–1.659, $\beta = 1.642$–1.660, $\gamma = 1.653$–1.680. The average length of sillimanite needles is 1.0 mm.

Diopside-scapolite granulite crops out only on Merry Hill, in the southeast corner of the north-central ninth of the quadrangle, surrounded by quartz monzonite. The outcrop of diopside-scapolite granulite is too small to show on the geologic map (Pl. 1). The rock is strongly banded with alternating layers of green diopside and flesh-colored to white scapolite. The layers range from 0.05 to 0.75 inch in thickness. Vuggy quartz veins cut irregularly across some specimens, and porphyroblasts of garnet are scattered through the rock. Plagioclase feldspars have apparently been altered by hydrothermal fluids to form the scapolite. A mode is given in Table 3, column 9.

In thin section, the rock has a granoblastic texture with euhedral to subhedral diopside and scapolite having mutual boundaries. Large subhedral grains of apatite are poikiloblastic in scapolite. The diopside contains approximately 40 per cent of the hedenbergite molecule. The scapolite is essentially a mizzonite with 60 per cent of the meionite molecule (Larsen and Berman, 1934, p. 245).

Actinolite granulite (Table 3, column 10) occurs around Watson Hill in the southwest part of Raymond township and is similar to rocks of this type found in the Berwick formation.

Thickness

The true thickness of the Littleton formation in the Mt. Pawtuckaway quadrangle cannot be calculated because of disruption by the plutons. Katz (1917) estimated it to be 550 (?) feet. In Maine, the breadth of outcrop of the Littleton formation is 3.5 miles. Assuming that

the ratio of breadth to thickness is the same as in the Eliot and Berwick formations, the total thickness of the Littleton formation would be 7,750 feet. The maximum thickness of the Gove member would be approximately 200 feet.

Method of Determining thickness of Strata

Determination of the thickness of the metamorphic rocks in the Mt. Pawtuckaway quadrangle presented a problem. Data on the minor folds are insufficient for a reasonable direct measurement of the thickness. A statistical approach was tried, based on the following formula proposed by Billings (oral communication) and to be discussed by him in a forthcoming publication:

$$t = b \cdot \frac{n_w}{n} \cdot r_w - b \cdot \frac{n_e}{n} \cdot r_e$$

t —thickness of formation
b —breadth of outcrop of formation measured perpendicular to strike
n_w—number of outcrops with top to west
n_e—number of outcrops with top to east
n —number of outcrops for which data on direction of top are available
r_w—average ratio of breadth of outcrop to thickness in observed outcrops with top to west
r_e—average ratio of breadth of outcrop to thickness in observed outcrops with top to east

Thus, if (1) a formation has a breadth of outcrop of 5000 feet, (2) 20 outcrops show top to west whereas 5 show top to east, (3) the average ratio of breadth of outcrop to thickness in outcrops with top to west is 0.6, and the average ratio of breadth of outcrop to thickness in outcrops with top to east is 0.4, the thickness is:

$$t = 5000 \times \frac{20}{25} \times 0.6 - 5000 \times \frac{5}{25} \times 0.4$$

$$t = 2400 - 400 = 2000$$

A ratio of thickness to breadth of outcrop of 0.5 is assumed. The ratio of beds with top to west to those with top to the east was taken from the ratio of drag folds that showed top to west and east. The proportion was 6 to 1. This formula gave the thicknesses presented in Figure 2 for each formation. The total thickness of the metamorphic rocks in the Mt. Pawtuckaway quadrangle based on this calculation is about 20,000 feet.

New Hampshire Magma Series

(Late Devonian?)

General Statement

Plutonic rocks of the New Hampshire magma series, ranging from gabbro to granite, intrude the metamorphic rocks. Rocks of this magma series occupy much of the northwestern half of the quadrangle and a small area in the southeastern corner. These rocks were called Winnipesaukee gneiss by Hitchcock (1877) and Roy and Freedman (1944). The name is discarded here inasmuch as the formation is unlike the typical Winnipesaukee gneiss (Billings, oral communication). The New Hampshire magma series is represented by five units: Exeter diorite, quartz diorite, quartz monzonite, binary granite, and microcline granite. Within each unit some of the rocks are foliated whereas others are massive. The Exeter diorite occupies the southeastern corner of the area, and is separated from the other plutonic rocks of the series by a broad belt of metamorphic rocks. Although no contacts separating the map-units within the large central mass of plutonic rocks have been observed in the field, the quartz diorite and the microcline granite may be separate intrusions. The binary granite and the quartz monzonite are apparently parts of a single intrusion. The microcline granite occurs in the southwestern part of the quadrangle. The quartz monzonite extends northeast from the microcline granite almost to the northern border of the area. The binary granite underlies the western and northwestern parts of the quadrangle. The quartz diorite occupies the northeastern part of the quadrangle. Where the granite is abundant it holds up the hills, but close to the metamorphic rocks the latter are more resistant.

Lithologic Characteristics

Exeter diorite.—C. H. Hitchcock (1877, part 2, p. 658–675) named the Exeter diorite. This

unit, exposed in the southeastern corner of the Mt. Pawtuckaway quadrangle, is a dark-greenish-gray to black, coarse-grained, equi-

Albite and Carlsbad twinning are common in the feldspar grains; sericitization is common. Rounded remnants of altered pyroxene are surrounded by

TABLE 4. APPROXIMATE MODES—NEW HAMPSHIRE MAGMA SERIES

	1	2	3	4	5
Number of thin sections	2	4	5	4	1
Quartz	2	16	23	24	40
Orthoclase		tr	30	17	
Microcline				34	52
Plagioclase	47	62	30	16	
Amphibole	5	4	3		
Pyroxene	28				
Uralite	15				
Biotite	2	13	11	4	3
Muscovite			tr	4	5
Chlorite	tr	tr	1	1	tr
Calcite		1			
Leucoxene			tr		
Sericite	tr		tr	tr	
Magnetite	tr	1	1	tr	
Ilmenite		tr	tr	tr	
Pyrite		tr	tr		
Sphene		1			
Garnet				tr	
Tourmaline					tr
Apatite	1	1	1	tr	
Zircon		tr	tr	tr	tr
Range in anorthite content in plagioclase	35–58	28–38	24–27	25–28	
Range of grain size in mm	0.5–5.0	0.05–3.0	0.5–10.0	0.5–10.0	0.05–2.0

1. Exeter diorite.
2. Quartz diorite.
3. Quartz monzonite.
4. Binary granite.
5. Microcline granite.

granular to porphyritic rock (Table 4, column 1) from a quartz diorite to gabbro. The grain size ranges from 0.5 to 5.0 mm. The only exposures in the quadrangle occur at the east edge of the quadrangle, 1 mile north of the South Side State Road. Other exposures occur in the Dover quadrangle to the east along the South Side State Road.

The chief minerals are feldspar and uralitized pyroxene. Amphibole, biotite, chlorite, and quartz are present in minor quantities, and subhedral apatite and magnetite are accessory. The feldspar, plagioclase ranging from andesine, An_{35}, to labradorite, An_{65}, comprises 30 to 65 per cent of the rock.

fibrous, light-green, faintly pleochroic uralite, in which magnetite grains are scattered. Biotite contains subhedral grains of zircon surrounded by pleochroic haloes and is partially altered to chlorite and magnetite. Interstitial quartz with undulatory extinction occurs in small grains with tiny dark inclusions.

Quartz diorite.—The quartz diorite (Table 4, column 2) is exposed in the northeastern part of the quadrangle; the best exposures occur on Mt. Misery and along the road around Ayers Pond. It is a dark-gray, fine-grained to medium-grained, granitoid, massive to faintly foliated rock with a gray feldspathic groundmass and scattered grains of amphibole

and biotite. The grain size ranges from 0.05 to 3.0 mm.

Essential minerals are feldspar, quartz, and biotite with a little amphibole and chlorite. Accessory minerals are euhedral to subhedral apatite, magnetite, sphene, pyrite, calcite and a few sillimanite needles. The texture is typically hypidiomorphic granular. Plagioclase, the dominant feldspar, comprises 50–80 per cent of the rock. It ranges from oligoclase, An_{28}, to andesine, An_{38}. Albite and Carlsbad twinning are common and the mineral is slightly zoned and sericitized. Quartz occurs interstitially, poikilitic in feldspar, and in vermicular and myrmekitic intergrowths. Biotite, which is altering to green chlorite and magnetite, is embayed by quartz and feldspar. Pleochroic haloes around zircon grains embedded in biotite are common. The γ index of the biotite ranges from 1.618–1.652, which suggests a biotite containing from 35 to 60 per cent of the annite-siderophyllite molecule. The amphibole is common hornblende; it is pleochroic with X = pale greenish yellow, Y = pale green, Z = leek green, $Z \wedge c = 20°$ to $40°$, $2V$ = large, negative, $\alpha = 1.658$, $\beta = 1.670$, $\gamma = 1.680$. The hornblende is partially altered to chlorite and magnetite.

Quartz monzonite. — Best exposures of the quartz monzonite (Table 4, column 3) occur on Adams Hill in Deerfield and on Tavern Hill in Nottingham. Good exposures were observed at the top of the large island in Pawtuckaway Pond, along the road south of the Pawtuckaway Mountains, and along the road to Deer Pond south of the Pawtuckaway Mountains. It is a light gray, medium-grained to coarse-grained, granitoid, massive to faintly foliated rock.

This body ranges from granite to granodiorite but quartz monzonite predominates. They are light-gray to dark-gray, medium-grained to coarse-grained, massive to foliated rocks.

Essential minerals are feldspar, quartz, and biotite; some specimens contain amphibole and chlorite. Apatite, magnetite, sphene, zircon, ilmenite, and sericite are accessory. The feldspar consists dominantly of plagioclase, orthoclase, and microcline. The plagioclase, which makes up from 0 to 40 per cent of the rocks, ranges from oligoclase, An_{25}, to andesine, An_{33}. Albite and Carlsbad twinning are common. Plagioclase that is poikilitic in large orthoclase grains is more sericitized than the orthoclase. Orthoclase ranges from 10 to 40 per cent of the rock. Perthite occurs in micrographic intergrowths with quartz in some rock. Quartz, which occurs interstitially and is poikilitic in orthoclase, makes up 5 to 40 per cent of the rock. Biotite, 5 to 20 per cent of the rock, alters to chlorite, magnetite and ilmenite. Apatite and ilmenite, which is altering to leucoxene, are poikilitic in biotite. Pleochroic haloes around zircon grains embedded in biotite are common. The γ index of the biotite ranges from 1.622 to 1.662, which suggests a biotite ranging from 38 to 75 per cent of the annite-siderophyllite molecule. Amphibole comprises 15 per cent of one specimen of granodiorite. Its optical properties are as follows: pleochroism, X = pale greenish-yellow, Y = green, Z = olive green, $\alpha = 1.661$, $\beta = 1.675$, $\gamma = 1.682$, $2V = 70°$, negative, $r>v$, $Z \wedge c = 24°$. A mineral with these properties is essentially a common hornblende. It is altered slightly to chlorite.

Binary granite.—Rocks assigned to the binary granite (Table 4, column 4) are chiefly granite and some quartz monzonite which contain both biotite and muscovite. They are white to light-gray, medium-grained to coarse-grained, massive to faintly foliated rocks with abundant feldspar and quartz. The texture is typically hypidiomorphic granular, although a few specimens show cataclastic textures. The grain size ranges from 0.5 to 10.0 mm.

The best exposures of the binary granite occur at the intersection of roads south of Northwood Narrows, along the wood road northeast of Northwood Narrows, and on the Central State Road, half a mile east of the intersection of roads south of Northwood Narrows near the northwest corner of the quadrangle. Old quarrying operations on good exposures were observed 1.25 miles west of the northern end of Nippo Pond in Barrington township.

Essential minerals are feldspar and quartz with small quantities of biotite, muscovite, and chlorite. Accessory minerals are apatite, magnetite, garnet, and tourmaline. The tourmaline has an ω index of 1.660, which indicates 60 per cent of the schorlite molecule. The feldspars are plagioclase, orthoclase, perthite, and microcline. The plagioclase is oligoclase, An_{24} to An_{27}, and makes up from 5 to 45 per cent of the rock. The plagioclase, some of which has reverse zoning, is sericitized. Orthoclase occurs in some grains as a vermicular intergrowth with quartz. A trace of perthite was observed in a few specimens. Microcline composes approximately 70

per cent of a few of the rocks. Quartz, 20 to 30 per cent of the rock, is interstitial to feldspar, poikilitic in feldspar, and occurs in vermicular intergrowths. Biotite is partially altered to chlorite and magnetite. Numerous zircon grains surrounded by pleochroic haloes are imbedded in the biotite in which $\gamma = 1.633$. This is essentially a biotite containing approximately 50 per cent of the annite-siderophyllite molecule. Muscovite is formed in cracks in the feldspar; $2V = 36°$, and $\beta = 1.586$, suggesting approximately 64 per cent of the potash, 34 per cent of the magnesium and 2 per cent of the ferric iron muscovite molecules.

Microcline granite.—Best exposures of the microcline granite occur on the South Side State Road at the Raymond-Candia town line, on the hill north of Bean Island, at Dumplingtown Hill, southeast of East Candia, on Robinson Hill in northwestern Raymond township, and on the hill half a mile east of B. M. 227 on South Side State Road in Raymond. The microcline granite is a light-gray to pink, massive to foliated, medium-grained to coarse-grained rock, with abundant pink microcline. It is typically granitoid although cataclastic in a few specimens. One thin section of the microcline granite was studied (Table 4, column 5).

Microcline and quartz constitute the essential minerals; some biotite and plagioclase are also present. Accessory minerals are garnet, tourmaline, and apatite. The plagioclase is oligoclase, An_{25}. Quartz is inequigranular and has undulatory extinction. The grain size ranges from 0.05 to 2.0 mm. Muscovite and biotite are slightly chloritized and contain grains of zircon surrounded by pleochroic haloes.

Pegmatite.—Numerous pegmatite dikes crop out in the Mt. Pawtuckaway quadrangle, although in only one area are they being exploited for commercial feldspar. They are most abundant in the northern part of the quadrangle. There are also pegmatites in the southwestern corner of the area where the Smith Mine is being worked. The exposures of the pegmatites range from short, irregular layers in granite to dikes 450 feet long and 50 feet wide. Feldspar, commonly perthite, is the most abundant mineral with lesser amounts of muscovite, biotite, and quartz. The following minerals have been observed by the writer or reported by others: garnet, tourmaline, beryl, pyrite, sphalerite, columbite, triphyllite, löllingite, and the decomposition products of uranium, gummite, autunite, and uranophane.

Most of the pegmatites occur in the plutonic rocks, but they seem to be best developed not far from the contacts of the metamorphic rocks. Inclusions of the country rock in the granitic rock form the walls of the pegmatites in a number of places. In the metamorphic rocks they are most abundant in the Littleton formation, but there are a few pegmatite dikes in the Berwick formation.

White Mountain Magma Series

(Mississippian?)

General Statement

The Mississippian (?) White Mountain magma series is represented by several igneous bodies in the Pawtuckaway Mountains (Pls. 1, 2) in Deerfield and Nottingham, by three small bodies in the vicinity of Little Rattlesnake Hill (Pl. 1) near the southern edge of the quadrangle in Raymond township and by numerous dikes.

Pawtuckaway Mountains

Roy and Freedman (1944) described the petrology of the Pawtuckaway Mountains, but further information on the structure has been derived from recent aerial photographs. The rocks in the Pawtuckaway Mountains form an alkalic complex similar to other igneous rocks of the White Mountain magma series in New England and Canada. The rocks are porphyritic gabbro, augite diorite, hornblende diorite, diabase, foliated hornblende diorite, hornblende-biotite diorite, porphyritic diorite, fine-grained monzonite, and coarse-grained monzonite (Pl. 1).

The dominant rocks are diorites and monzonites. The monzonites underlie the higher portions of the area, whereas the diorites underlie the valleys. Lamprophyre, monzonite, and aplite dikes are abundant.

Little Rattlesnake Hill Area

Introduction.—Three small bodies of rocks belonging to the White Mountain magma series

crop out near the south boundary of the quadrangle near Little Rattlesnake Hill in Raymond. One body is on Little Rattlesnake Hill; a second 0.5 mile southeast of Little Rattlesnake Hill; a third is on a hill 0.5 mile northeast of Little Rattlesnake Hill. The rocks are classified into: (1) gabbro and monzonite, and (2) volcanic rocks. The coarse-grained gabbro and monzonite occur as cores of small bodies partially surrounded by the volcanic rocks.

TABLE 5. APPROXIMATE MODES—WHITE MOUNTAIN MAGMA SERIES
Little Rattlesnake Hill Area

	1	2	3	4	5	6	7
Number of thin sections	1	1	1	1	1	1	1
Quartz			1	10	30	tr	
Orthoclase			40		25		
Microperthite			45	84			
Plagioclase	70	65	5		25	93	70
Biotite	8	5	1	tr	10	5	20
Amphibole						tr	
Augite	10	15		5	8		
Hypersthene		5	8				
Olivine	7						
Pyrite		tr		tr	tr	1	
Magnetite	5	10	tr		tr	1	10
Apatite	tr	tr	tr		tr		
Zircon			tr	tr	tr		
Anorthite content, range given where zoned	65–77	52	26		28–45	03–38	25
Range of grain size in mm.	2.0–5.0	0.5–5.0	2.0–7.0	0.03–2.0	0.01–1.0	0.01–2.0	0.01–1.0

1. Gabbro.
2. Hypersthene gabbro.
3. Syenite phase of hypersthene monzonite.
4. Rhyolite.
5. Quartz latite.
6. Keratophyre.
7. Andesite.

Gabbro and monzonite.—The bodies mapped as gabbro and monzonite (Pl. 1) consist of several kinds of rock including gabbro, hypersthene gabbro, and hypersthene monzonite. In hand specimen, coarse-grained laths of plagioclase show albite twinning and an iridescent luster typical of labradorite.

Microscopic study reveals that the gabbro (Table 5, column 1), collected 0.5 mile southeast of Little Rattlesnake Hill, has an ophitic texture and is made up chiefly of plagioclase, olivine, pyroxene, and biotite. Accessory minerals are apatite and magnetite. The plagioclase is labradorite-bytownite, An_{63-77}, which has both normal and reverse zoning. It occurs in long laths, showing albite and Carlsbad twinning, and in small euhedral crystals poikilitic in the pyroxene. The pyroxene is pink, titaniferous augite. Olivine and feldspar are poikilitic in it and large, subhedral magnetite grains are an alteration product. The olivine is chrysolite, containing approximately 30 per cent of the fayalite molecule. It is altered along cracks and around the rims to a fibrous, greenish serpentine. A few flakes of biotite are almost completely altered to magnetite, whereas many flakes are unaltered.

The essential minerals in the hypersthene gabbro also from the body 0.5 mi. southeast at Little Rattlesnake Hill, are plagioclase, augite, hypersthene, and biotite. Apatite and pyrite are the accessory minerals (Table 5, column 2). The plagioclase, labradorite, An_{52} is locally sericitized. Albite and Carlsbad twinning are common. Augite and hypersthene are partially altered to biotite and chlorite. Apatite occurs poikilitically in both. Biotite in subhedral grains is partially to almost completely altered to anhedral magnetite.

Hypersthene monzonite is predominant on Little Rattlesnake Hill. Although most specimens are similar to the coarse monzonite of the Pawtuckaway Mountains, the specimen selected for thin section was a syenite. This rock is composed chiefly of antiperthite, orthoclase, oligoclase, and hypersthene, and very minor amounts of biotite and quartz. The accessory minerals are magnetite, apatite, and zircon (Table 5, column 3). Grain size ranges from 2.0 to 7.0 mm. The plagioclase is normally zoned oligoclase which averages 26 per cent anorthite. The orthoclase occurs as irregular grains. Additional orthoclase is present in microperthitic intergrowths, which comprise half of the feldspar of the rock. Hypersthene is euhedral to subhedral and is partially altered to hornblende. Quartz is interstitial and a few grains of biotite are scattered throughout the rock. A porphyritic, fine-grained rock containing phenocrysts of feldspar, similar to the feldspar in the coarse-grained rocks, crops out at the northwest end of Little Rattlesnake Hill. It is apparently a contact phase of the coarse-grained hypersthene monzonite core.

Volcanic rocks.—Good exposures of the volcanic rocks are found in the vicinity of all three bodies. They are light-bluish-gray to dark-gray, massive, porphyritic, and brecciated rocks. The volcanic rocks are variable in composition and include rhyolite, quartz latite, keratophyre, andesite, and breccia.

The rhyolite, which crops out on Little Rattlesnake Hill, is bluish-gray on a fresh surface but weathers white. Under the microscope the texture is typically hypidiomorphic granular. Grain size ranges from 0.03 to 2.0 mm. Essential minerals are feldspar and quartz, with a little augite and biotite. Accessory minerals are zircon and pyrite (Table 5, column 4). The feldspar is microperthite which comprises most of the rock.

Quartz latite, which crops out 0.5 mile southeast of Little Rattlesnake Hill, is a bluish-gray rock with a hypidiomorphic inequigranular texture. Essential minerals are feldspar and quartz with a little biotite and augite. Accessory minerals are magnetite, apatite, pyrite, and zircon (Table 5, column 5). Plagioclase, oligoclase-andesine, is zoned with an anorthite content ranging from 28 to 45 per cent. Orthoclase occurs interstitially. The augite has been slightly chloritized. The average grain size of the rock ranges from 0.01 to 1.0 mm.

The keratophyre crops out 0.5 mile southeast of Little Rattlesnake Hill. It is blue-gray with small feldspar phenocrysts and abundant anhedral pyrite. Under the microscope, the texture is hypidiomorphic porphyritic. Grain size ranges from 0.01 to 2.0 mm. Feldspar is the dominant mineral and a little biotite, amphibole, and quartz occur with it. Accessory minerals are pyrite, magnetite, and apatite (Table 5, column 6). The feldspar makes up 93 per cent of the rock, of which 10 per cent are phenocrysts composed of slightly zoned, calcic oligoclase, dominantly An_{28}. The remainder of the feldspar, which is albite, An_3, makes up the groundmass. Scattered flakes of biotite form both phenocrysts and part of the groundmass. A trace of brown basaltic hornblende in which $Z \wedge c = 7°$ is present. Quartz and pyrite occur intergrown in round knots.

The andesite is a dark-gray to black, dense rock which crops out on a hill 0.5 mile northeast of Little Rattlesnake Hill. Grain size ranges from 0.01 to 1.0 mm. Feldspar and biotite are the chief minerals; magnetite constitutes 10 per cent of the rock (Table 5, column 7). The feldspar, oligoclase, An_{21}, is sericitized and occurs in tiny laths. The biotite is highly chloritized and occurs interstitially. Magnetite, locally altered to limonite, occurs in scattered grains.

Breccia occurs in one outcrop on the top of the hill 0.5 mile northeast of Little Rattlesnake Hill and comprises much of the body 0.5 mile southeast of Little Rattlesnake Hill. The breccia on the hill to the northeast is made up of angular fragments, about 10 mm. across, of dense andesite, coarser-grained biotite schist, and fine-grained rhyolite containing abundant pink feldspar and some quartz. The rhyolite also binds the breccia fragments together.

The breccia at the locality southeast of Little Rattlesnake Hill is the chief rock outside of a dominantly gabbroic core. Inclusions of the wall-rock quartzite and quartz-biotite schist, some of which have altered rims, occur in the blue-gray, fine-grained quartz latite on the low ridge, 300 feet west of the two houses. Granite and pegmatite fragments are locally numerous. Since neither granite nor pegmatite crops out in the vicinity, these have probably been brought up from below. Average size of the fragments is about 25 mm. The breccia is also abundant northwest of the gabbro core. Here the breccia consists of large fragments, as much as 2 feet across of quartz-biotite schist, quartzite, and fine-grained and coarse-grained fragments of rock composing the core and the rocks around it. The matrix of the breccia is dominantly fine-grained material.

Dike and Sill Rocks

General Statement

The dikes and sills in the Mt. Pawtuckaway quadrangle may be assigned to two groups:

1) pegmatites related to the New Hampshire magma series; and 2) trachyte, monzonite, aplite, and mafic rocks, most of which are related to the White Mountain magma series. The dikes in the Pawtuckaway Mountains,

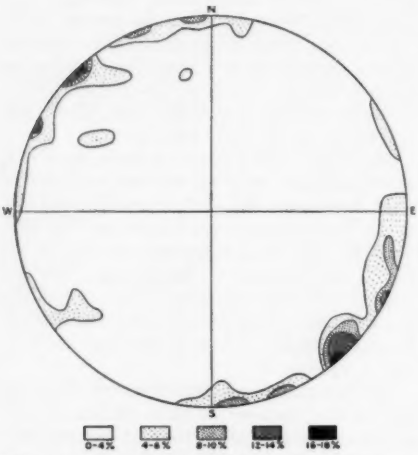


FIGURE 4.—CONTOUR DIAGRAM OF FIFTY-FIVE DIKES AND SILLS
(Poles of perpendiculars plotted on upper hemisphere of equal area net)

chiefly monzonite, aplite, and lamprophyres, have been described in a previous paper (Roy and Freedman, 1944, p. 912–913).

Structure

Fifty-five dikes and sills are sufficiently well exposed to measure their attitude; 35 of them are pegmatites belonging to the New Hampshire magma series, and 20 are trachytes and mafic rocks related to the White Mountain magma series. Dikes in the Pawtuckaway Mountains are excluded; their orientation is controlled locally and would obscure the regional relationship. The attitude of these 55 dikes and sills (Fig. 4) is similar to that of the joints (Figs. 8, 9). The diagram (Fig. 4) is a plot of the poles of perpendiculars to the dikes and is plotted on the upper hemisphere of an equal area net as suggested by Billings (1942, p. 116–121). Figure 4 indicates that most of the dikes and sills strike northeast and dip either vertically or steeply northwest or southeast. The dikes and sills have probably been injected parallel to the bedding of the metamorphic rocks or along joints in the igneous rocks. A small concentration at the north and south poles of the diagram indicates a set of dikes and sills striking approximately 45° to the dominant regional trend.

The trachyte and mafic dikes and sills range from 1 to 12 feet thick, but the average is approximately 2 feet. The pegmatite dikes and sills, intimately associated with rocks of the New Hampshire magma series and never found with younger rocks, belong to the New Hampshire magma series. All others, except the diabase dikes and sills, on the basis of their mineral composition are co-magmatic with the White Mountain magma series. The diabase dikes are possibly of Triassic age.

Petrography

From thin sections made from 12 trachyte and mafic dike and sill rocks the following types have been recognized: trachyte, hornblende diorite, augite diorite, camptonite, ouachitite (Grout, 1932, p. 124), and diabase (Table 6).

Two trachyte dikes were observed. One, exposed at the sharp bend in the road 900 feet due east of the house shown on Adams Hill in Deerfield, strikes N.45°E. A second, exposed on the summit of hill elevation 383 in the northeastern part of Raymond township, strikes N.65°E. Both are 18 inches thick. Specimens of each were studied microscopically. The first is a white, aphanitic, iron-stained rock; the other is dark-greenish-gray and aphanitic. They are composed dominantly of orthoclase with some chlorite and limonite; accessory minerals are biotite, hornblende, and calcite (Table 6, column 1). The rocks have a felted texture and grain size ranges from 0.01 to 0.5 mm.

The specimen of hornblende diorite comes from an outcrop 1200 feet northwest of Nottingham Square on the northeast side of the road. The dike strikes N.60°E. and is 3.5 feet thick. The hand specimen is a dark-gray, medium-grained porphyritic rock with feldspar phenocyrsts and a felted groundmass. Essential minerals determined in thin section are plagioclase and green hornblende; accessory minerals are ilmenite, magnetite, and pyrite (Table 6, column 2). The plagioclase is andesine (An_{40}) and comprises more than half of the rock. Albite and Carlsbad twinning were observed and most of the grains are partly altered to calcite. The green hornblende has apparently been derived from augite and is partially altered to chlorite and magnetite.

Two dikes of augite diorite were observed. One, exposed 1 mile due north of the village of Nottingham, strikes N.10°E. The second, exposed along the road 0.9 mile S.50°E. of West Epping, strikes due east. At the first locality, two sills, 1 foot and 2 feet thick, are composed of a dark-gray, fine-grained

rock which contains small cubes of pyrite. At the second locality, where the dike is 20 feet thick, the rock is dark-gray with phenocrysts of hornblende. The essential minerals are feldspar, pyroxene, and chlorite, with a little biotite. Accessory minerals are strikes N.10°W. and is 6 feet thick. Both dikes are black, fine-grained porphyritic rocks, with phenocrysts of barkevikite and feldspar in a felted groundmass. Under the microscope, dominant minerals are barkevikite, feldspar, and pyroxene. Accessory

TABLE 6. APPROXIMATE MODES—DIKE AND SILL ROCKS

	1	2	3	4	5	6
Number of thin sections	2	1	3	2	1	3
Quartz		tr				
Orthoclase	72					
Plagioclase		55	40	57	73	45
Hornblende	tr	36		23		
Pyroxene			13	3	10	31
Biotite	1		3		15	tr
Muscovite						
Sericite	tr		tr			tr
Chlorite	20	tr	38	tr		13
Epidote						7
Calcite	tr	2	tr	8		tr
Sphene			5			
Magnetite		2	tr	5		4
Limonite	7					
Ilmenite		5	tr	tr		
Apatite			tr	1		
Pyrite		tr	1	3		
Kaolin				tr		
Range in anorthite content in plagioclase		45	43–45	45–55	60	50–70
Range of grain size in mm.	0.01–0.5	1.0–5.0	0.1–10.0	0.01–5.0	0.05–1.0	0.01–1.0

1. Trachyte.
2. Hornblende diorite.
3. Augite diorite.
4. Camptonite.
5. Ouachitite.
6. Diabase.

apatite, magnetite, ilmenite, pyrite, and sphene (Table 6, column 3). The feldspar, an andesine which contains 43 to 45 per cent anorthite, ranges from 15 to 55 per cent of the rock. The feldspar grains, which occur as laths and microlites, are slightly sericitized. The pyroxene is augite that ranges from 3 to 20 per cent of the rock. Chlorite is colorless to green and constitutes 20 to 50 per cent of the rock. Grain size ranges from 0.1 to 10.0 mm.

Two camptonites were observed. One crops out in a gravel pit near the bridge over the North River 2 miles S.40°E. of West Nottingham. A second is exposed on an old road 1300 feet southeast of Northwood Narrows. At the first locality, the dike strikes N.10°E. and is 4 feet thick; at the second, the dike minerals, which range from 5 to 10 per cent, are magnetite, apatite, ilmenite, calcite, and pyrite (Table 6, column 4). The feldspar, ranges from andesine, An_{45}, to labradorite, An_{55}. Albite and Carlsbad twinning are common. The barkevikite has the following optical properties: X = pale yellow, Y = light brown, Z = brown, $Z \wedge c = 12°$ to 16°, $\alpha = 1.671$, $\beta = 1.688$, $\gamma = 1.702$, negative, $2V = 80°$. Pyroxene is a titaniferous augite showing a faint pink pleochroism and kelyphitic rims of chlorite and amphibole.

The ouachitite occurs as a dike 10 feet thick 0.5 mile S.50°E. of Little Rattlesnake Hill in Raymond. It is a black, porphyritic rock with an aphanitic groundmass and numerous, small feldspar phenocrysts. The essential minerals are feldspar, biotite,

and pyroxene. Accessory minerals are pyrite and magnetite (Table 6, column 5). The feldspar is labradorite, An_{60}, and comprises about three-fourths of the rock. The grains have indistinct zoning and

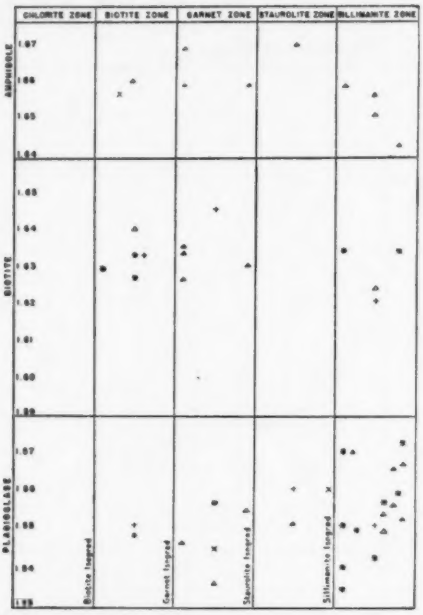


FIGURE 5.—RELATIONSHIP OF MINERAL INDICES TO METAMORPHIC ZONES

Vertical scale is gamma index. Triangle = lime-silicate rocks; cross = quartzite and quartz-mica schist; plus = mica-quartz schist; asterisk = mica schist.

Carlsbad twins are abundant. Biotite is chloritized. The pyroxene, augite, occurs in anhedral grains which are partially altered to biotite and chlorite.

Three diabase dikes were observed. One, exposed along the highway in the northeastern part of Nottingham 1 mile southeast of Merry Hill, strikes N.40°E. and is 60 feet thick. A second, exposed south of the Central State Road 2 miles S.50°E. of Merry Hill, strikes N.60°E. A third dike, 100 feet thick, is exposed in a railroad cut 0.6 mile S.60°E. of Onway in Raymond and strikes N.25°E. The first and third dikes are dark-gray to black, phanerocrystalline rocks and the second dike is black and aphanitic. They have an ophitic texture and are composed dominantly of feldspar, pyroxene, and chlorite; epidote is abundant in one specimen. Accessory minerals are sphene and magnetite (Table 6, column 6). The feldspar, which constitutes 30 to 54 per cent of the rock, is labradorite, An_{50} to An_{70}. It occurs in unoriented laths which show albite and Carlsbad twins. The pyroxenes, which are augite altered to green chlorite, constitute 20 to 44 per cent of the rocks.

COMPARATIVE MINERALOGY

Optical data were obtained on the essential minerals in most of the rocks for which thin sections were available. The chemical composition of the minerals which belong to isomorphous series was determined optically where the relationship between optical data and chemical composition is known.

To determine whether the isomorphous minerals showed regular changes in chemical composition with increase of metamorphism, the gamma indices of plagioclase, biotite, and amphibole in the metamorphic rocks were plotted by rock type in the successive metamorphic zones (Fig. 5). Incomplete data make the results tentative, but suggest that there is little progressive change in the successive metamorphic zones. The concentration of plagioclase in the sillimanite zone reflects either an easier separation of feldspar from that zone, or closer proximity of that zone to magma with consequent enrichment of feldspar-rich fluids.

Comparison of the mineralogical data of the White Mountain and New Hampshire magma series, lime-silicate rocks, paraschists, and dike and sill rocks (Fig. 6) is more illuminating. The γ index is the abscissa; the per cent occurrence of the γ index of a mineral in a range of 0.005 is the ordinate. For example, in the plagioclase diagram, the curve for the New Hampshire magma series reaches 37 per cent at 1.553. This indicates that of all the determinations made, 37 per cent fall between 1.550 and 1.555. The point within each 0.005 range was placed at the highest percentage within that range.

The curves in the plagioclase, biotite, amphibole, and pyroxene diagrams are correlated with scales which indicate the chemical composition. The curves show the range and variation of these minerals in each lithologic unit, and their variation between lithologic units.

The plagioclase in the rocks of the Mt. Pawtuckaway quadrangle shows a wide range from sodic albite in some rocks of the White Mountain magma series to sodic bytownite in some of the dike rocks. The plagioclase in the New Hampshire magma series, the White Mountain magma series (exclusive of dikes and sills), and the lime-silicate rocks is chiefly oligoclase and andesine. The plagioclase

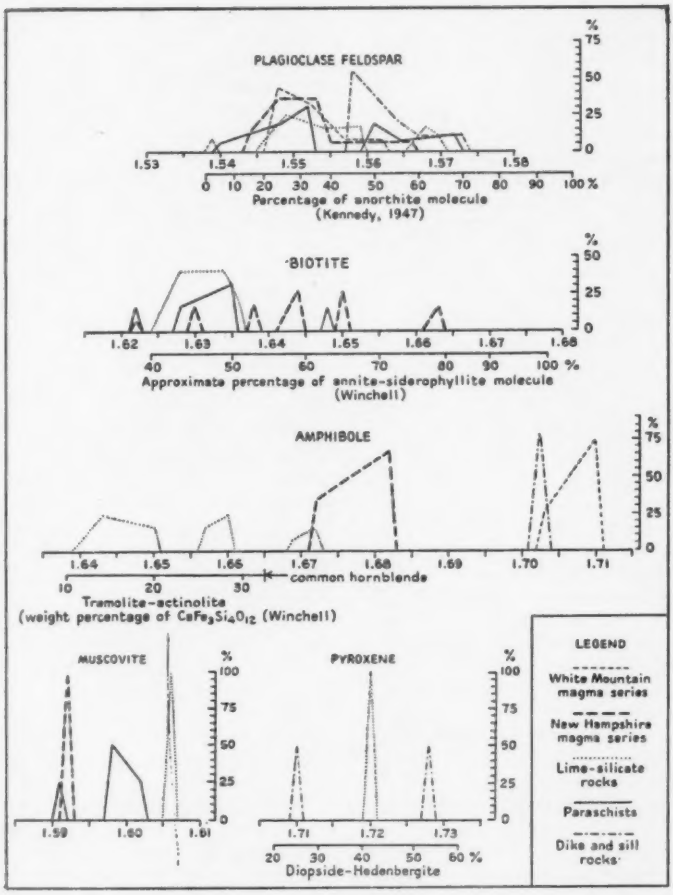


FIGURE 6.—COMPARATIVE MINERALOGY

γ index is the abscissa; the per cent occurrence of the γ index of a mineral in a range of 0.005 is the ordinate.

in the dike and sill rocks is andesine and labradorite consistent with their mafic aspect. Some of the plagioclase of the lime-silicate rocks is labradorite; most of it is oligoclase and andesine. The plagioclase of the paraschists is chiefly albite, oligoclase, and sodic andesine; there is also considerable labradorite. Thus the contrast between the lime-silicate rocks and the paraschists is not very great, in contrast to the observations of Billings (1941, p. 903) and Kruger (1946, p. 199), who found that the plagioclase in the paraschists tended to be calcic andesine, labradorite, and bytownite.

The biotite in the paraschists of the Berwick and Littleton formations shows three scattered maxima of the γ index that fall at 1.622, 1.635, and 1.648. One maximum for the γ index of the biotite in the lime-silicate rocks extends from 1.627 and 1.634. Maxima for the New Hampshire magma series are also scattered and fall at 1.622, 1.630, 1.638, 1.644, 1.650 and 1.662, giving a wide distribution of small percentages.

The biotite in the lime-silicate rocks, although relatively rich in magnesium compared to other groups, is also higher in iron that that of similar rocks elsewhere in New Hampshire. However, Kruger (1946, p. 198) shows a range of 1.611 to 1.652 compared with 1.625 to 1.636 in the present study. Billings (1941) shows the biotite in the

New Hampshire magma series as ranging from 1.640 to 1.660. Approximately 50 per cent of the determinations of the Mt. Pawtuckaway quadrangle fall within that range.

Amphibole (Fig. 6) is abundant in the lime-silicate rocks, the White Mountain magma series, the dike and sill rocks, and locally in the New Hampshire magma series. In the lime-silicate rocks, approximately 50 per cent of the γ indices range between 1.640 and 1.650. Optical data indicate that the amphibole is actinolite with about 20 per cent of the ferrotremolite molecule ($CaFe_5Si_4O_{12}$). The amphibole with a submaximum at 1.672 is actinolitic hornblende.

The γ indices of the amphibole of the New Hampshire magma series are higher than those of the actinolite and range from 1.672 to a maximum at 1.682. Complete optical data indicate that the amphibole is common hornblende.

In the dike and sill rocks, the γ indices are about 1.703. Optical data show these to be barkevikite, a brown amphibole, similar to that at the type locality at Livermore Falls, Campton, New Hampshire (Billings, 1928).

The amphibole in the White Mountain magma series has the highest indices of all; these, ranging from 1.702 to 1.710, are essentialy basaltic hornblende (Roy and Freedman, 1944). They differ from the barkevikite in optical orientation—$Z \wedge c$ = 9° in the basaltic hornblende, and ranges from 12° to 16° in the barkevikite.

The γ index of muscovite ranges from 1.591 to 1.606 in the paraschists, lime-silicates and the New Hampshire magma series.

The maximum for the paraschists is 50 per cent where γ = 1.598. Submaxima occur at 1.591 and 1.602. Single determinations of muscovites of the lime-silicate rocks and the New Hampshire magma series show γ = 1.606 and 1.592 respectively. The data are insufficient for reliable comparative purposes.

These are dominantly potash muscovites low in ferric iron, with the iron content lowest in the New Hampshire magma series, higher in the paraschists, and still higher in the lime-silicate rocks. Magnesian muscovite, not indicated on this graph, comprises as much as 34 per cent of the muscovite. The $2V$ of the muscovites ranges from 36° to 38°. Kruger (1946) shows muscovite with a γ index ranging from approximately 1.599 to 1.616. These are slightly richer in ferric iron. In the Mt. Washington quadrangle (Billings, 1941) the γ index of the muscovite of the lime-silicate rocks of the Boott member of the Littleton formation reaches a maximum at 1.600 coinciding with that studied here.

The pyroxene, augite, is abundant only in the dike and sill rocks. The γ indices for the dike and sill rocks are variable. Two maxima occur at 1.710 and 1.728. The pyroxenes contain approximately 25 per cent and 53 per cent by weight of the hedenbergite molecule respectively. The latter is similar to a pyroxene in the Merrymeeting Lake area, New Hampshire (Quinn and Stewart, 1941).

Only one outcrop of lime-silicate rocks contains pyroxene. The γ index of this pyroxene in the lime-silicate rocks is 1.720. Complete optical data indicate a diopside containing 40 per cent by weight of the hedenbergite molecule. This is richer in iron than diopside from the Boott member of the Littleton formation (Billings, 1941).

Indices of refraction were not obtained on hypersthene or titaniferous augite in a few specimens of rock from the White Mountain magma series in the Little Rattlesnake Hill area. The hypersthene would fall at the lower end of the graph, and the titaniferous augite at the upper end.

Chlorite, most abundant in the Eliot formation, also occurs in significant amounts in the other metamorphic formations and in the igneous rocks. Most of the chlorites are penninite, ripidolite, or diabantite. The γ index of diabantite from one specimen is 1.620, and that of ripidolite from another specimen is 1.623. These data indicate chlorites with approximately 50 per cent of the amesite molecule. Penninite contains approximately 30 per cent of the amesite molecule.

Staurolite and sillimanite occur only in the Littleton formation. Since both minerals have constant optics, they have constant chemical composition.

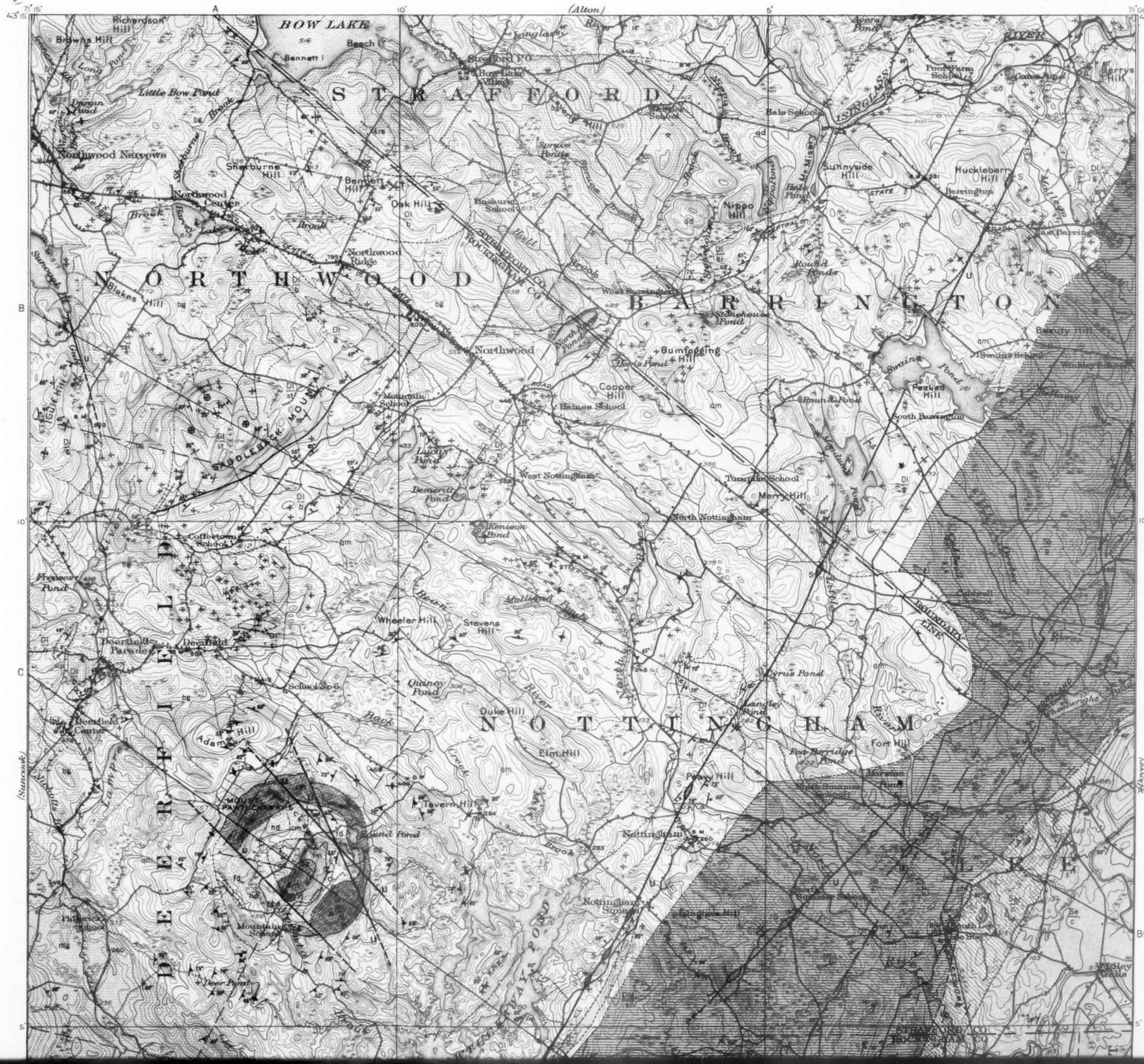
Optical data were obtained on garnets from schists of the Littleton formation, from pegmatite, and from a lime-silicate rock. The indices of the garnets from the schist and pegmatite are uniform with an approximate index of 1.807. Coupled with data on the specific gravities, which differ slightly, the data indicate that the garnets are almandite with some spessartite. These are similar to garnet reported by Billings (1941, p. 904).

The index of the garnet from the lime-silicate rock is 1.778. A reliable specific gravity was unobtainable, so an X-ray powder picture was taken to determine the unit cell. The garnet was determined to be 70 per cent pyrope + almandite and 30 per cent grossularite + andradite. This garnet has a lower index than garnet from lime-silicate rocks reported by Kruger (1946).

Potash feldspar occurs in four specimens of the paraschists. Orthoclase comprises from 1 to 10 per cent of three of these; the fourth contains 65 per cent microperthite. The paraschists are similar to rocks of the Mt. Washington area in their scarcity

NEW HAMPSHIRE STATE PLANNING
AND DEVELOPMENT COMMISSION AND HIGHWAY DEPARTMENT

NEW HAMPSHIRE
MT. PAWTUCKAWAY QUADRANGLE


(Gilmanton)
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71°15'
71°00'
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A
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STRAFFORD
NORTHWOOD
BARRINGTON
DERRY
NOTTINGHAM
LEE
BOW LAKE
Richardson Hill
Brown Hill
Bennett Hill
Little Bow Pond
Northwood Narrows
Sherburne Hill
Bennett Hill
Northwood Center
Oak Hill
Northwood Ridge
Blakes Hill
Northwood
Bumfagging Hill
Cooper Hill
Haines School
Mountain School
Saddleback Mountain
Coffeetown School
Wheeler Hill
Stevens Hill
Deerfield Parade
Deerfield
Adams Hill
Mount Pawtuckaway
Tavern Hill
Quaint Pond
Deer Pond
Duke Hill
Elm Hill
Nottingham
Nottingham Square
Nottingham Hill
West Nottingham
North Nottingham
Merry Hill
Tuxmire School
Kenison Pond
Quincy Pond
Cyrus Pond
Langley Pond
Red Barridge Pond
Round Pond
Swains Pond
Pawtuckaway Pond
Northwood Lake
Bow Lake Village
Bow Lake
Stuart P.O.
Barker School
Halls School
Sunnyside Hill
Huckleberry Hill
Barrington
East Barrington
South Barrington
Beauty Hill
Pearl Hill
Nippo Hill
Foss Farm School
Berrys Hill
Isinglass River
Swains School
Winkley Corners
Meaderboro
Round Pond
Caldwell School
Boundary Line
Fox Hill
Morrison Pond
Lee Traverse
Lee Circle
Wadley Falls
Pinkerton Academy
Deerfield Parade Center
Pinkerton Academy
Mountain School
Pleasant Pond
Deerfield Center
Northwood Narrows
Northwood Center
Northwood Ridge
Northwood Lake
Bow Lake Village
Stuart P.O.
Barker School
Halls School
Foss Farm School
Caldwell School
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Lee Traverse
Lee Circle
Wadley Falls
Pinkerton Academy
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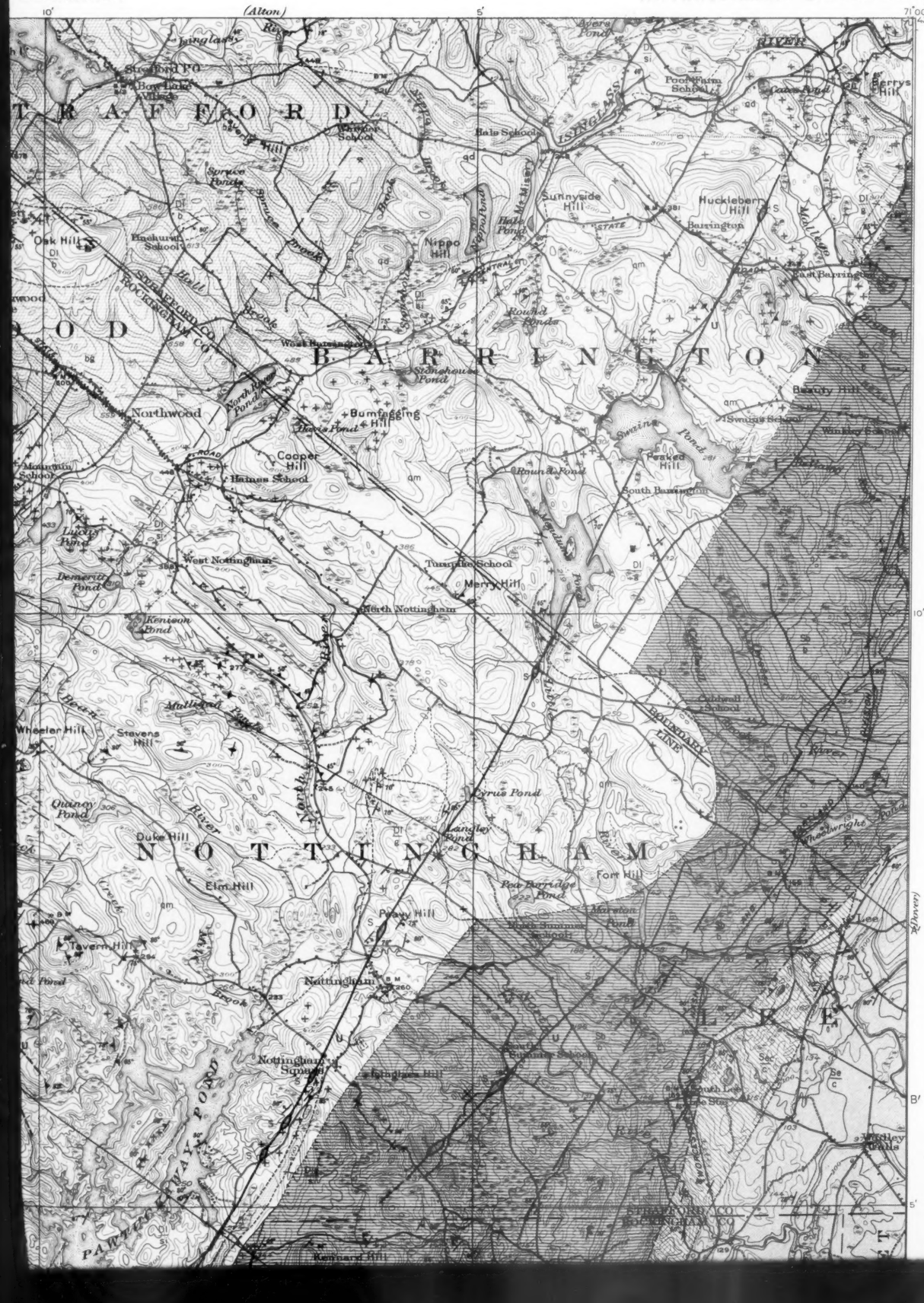
G
Y DEPARTMENT

NEW HAMPSHIRE
MT. PAWTUCKAWAY QUADRANGLE



LEGEND

METAMORPHIC AND IGNEOUS ROCKS

MISSISSIPPIAN?

White Mountain Magma Series

- gm — **Coarse-grained monzonite**
  (Coarse-grained, light-gray to dark-gray monzonite, composed chiefly of oligoclase-andesine, orthoclase, microcline, hornblende, biotite, and augite.)
- fm — **Fine-grained monzonite**
  (Fine-grained, dark-gray to blue-gray monzonite, composed chiefly of oligoclase, orthoclase, microcline, biotite, hornblende, and augite.)
- pd — **Porphyritic diorite**
  (Coarse-grained, porphyritic diorite, composed of hornblende phenocrysts 2 to 5 cm. long, set in a white granular groundmass composed of oligoclase, biotite, orthoclase, and augite.)
- hbd — **Hornblende-biotite diorite**
  (Medium-grained, gray diorite, composed of oligoclase, biotite, hornblende, orthoclase, and augite.)
- fd — **Foliated hornblende diorite and diabase**
  (Foliated hornblende diorite: foliated, coarse-grained, dark-gray diorite, composed of andesine-labradorite, hornblende, biotite, augite, and olivine. Diabase: coarse-grained, dark-gray diabase with ophitic texture, composed of andesine-labradorite, hornblende, biotite, augite, and olivine.)
- hd — **Hornblende diorite**
  (Coarse-grained, dark-gray to black diorite, composed of andesine, hornblende, biotite, augite, and olivine.)
- pg — **Porphyritic gabbro and augite diorite**
  (Porphyritic gabbro: medium-grained, black porphyritic gabbro with phenocrysts of augite up to 1.5 cm. in diameter set in a groundmass composed of labradorite, augite, hornblende, biotite, and olivine. Augite diorite: medium-grained, black diorite, composed of zoned plagioclase ranging from oligoclase to andesine, augite, hornblende, and biotite.)
- v — **Volcanic rocks**
  (Rhyolite: bluish-gray, dense rhyolite that weathers white. Quartz latite: bluish-gray quartz latite with phenocrysts of orthoclase and plagioclase. Keratophyre: bluish-gray keratophyre with phenocrysts of oligoclase up to 0.3 cm. long. Andesi dark-gray to blue andesite. Breccia: breccia composed of angular fragments, 1.0 to 2.5 cm. in diameter, composed of andesite, rhyolite, and schist set in a rhyolite matrix.)
- g — **Gabbro and monzonite**
  (Gabbro: coarse-grained gabbro, composed of white laths of labradorite-bytownite, with interstitial augite, olivine, and magnetite. Hypersthene gabbro: coarse-grained hypersthene gabbro, composed of labradorite, augite, magnetite, hypersthene, and biotite. Monzonite: coarse-grained hypersthene monzonite, composed of orthoclase, microperthite, oligoclase, and hypersthene.)
- \+ — **Pegmatite**
  (Composed chiefly of perthite with some muscovite, biotite, and quartz.)

LOWER DEVONIAN

SILURIAN?

- Dl / Dlg / Dl — **Littleton formation**
  (Zone b: gray phyllite and quartzite, with some mica schist and mica-quartz schist. Zone g: mica schist, mica-garnet schist, and quartzite. Zone st: mica schist, mica-garnet schist, mica-garnet-staurolite schist, and quartzite. Zone si: mica schist, mica-garnet schist, mica-sillimanite schist, and quartzite. Dlg, Gove member: chiefly silvery mica schist, some mica-staurolite schist, mica-sillimanite schist, and quartzite.)
- Su — **Berwick formation**
  (Zone b: black, gray and green phyllite with biotite occurring both as porphyroblasts and in groundmass; quartz-mica schist; quartzite; actinolite-quartz granulite; actinolite granulite; biotite-actinolite schist; actinolite amphibolite. Zone g: similar to rocks in Zone b, but some of them contain garnet.)
- Sec / Se — **Eliot formation**
  (Zone c: dark-gray, light-gray and green phyllite; quartz-sericite schist; quartzite; calcareous phyllite. Zone b: dark-gray, light-gray, and green phyllite with biotite porphyroblasts; quartz-mica schist with biotite porphyroblasts; quartzite; actinolite-biotite granulite. Sec, Calef member; chiefly black phyllite; some green quartz-chlorite phyllite.)

METAMORPHIC ZONES

Shown below formation symbol thus: Dl/g

- c - chlorite zone
- b - biotite zone
- g - garnet zone
- st - staurolite zone
- si - sillimanite zone

CONTACTS

- Accurate
- Approximate and diagrammatic due to poor exposures
- Indefinite as sharp contact is lacking

SPECIAL SYMBOLS

- Strike and dip of bedding
- Strike of vertical beds
- Horizontal bedding
- Strike and dip of foliation of igneous and metamorphic rocks

G
Y DEPARTMENT

NEW HAMPSHIRE
MT. PAWTUCKAWAY QUADRANGLE

LEGEND

METAMORPHIC AND IGNEOUS ROCKS

MISSISSIPPIAN? — *White Mountain Magma Series*

- cm — **Coarse-grained monzonite**
  (Coarse-grained, light-gray to dark-gray monzonite, composed chiefly of oligoclase-andesine, orthoclase, microcline, hornblende, biotite, and augite.)
- fm — **Fine-grained monzonite**
  (Fine-grained, dark-gray to blue-gray monzonite, composed chiefly of oligoclase, orthoclase, microcline, biotite, hornblende, and augite.)
- pd — **Porphyritic diorite**
  (Coarse-grained, porphyritic diorite, composed of hornblende phenocrysts 2 to 5 cm. long, set in a white granular groundmass composed of oligoclase, biotite, orthoclase, and augite.)
- hbd — **Hornblende-biotite diorite**
  (Medium-grained, gray diorite, composed of oligoclase, biotite, hornblende, orthoclase, and augite.)
- fd — **Foliated hornblende diorite and diabase**
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  (Coarse-grained, dark-gray to black diorite, composed of andesine, hornblende, biotite, augite, and olivine.)
- pg — **Porphyritic gabbro and augite diorite**
  (Porphyritic gabbro: medium-grained, black porphyritic gabbro with phenocrysts of augite up to 1.5 cm. in diameter set in a groundmass composed of labradorite, augite, hornblende, biotite, and olivine. Augite diorite: medium-grained, black diorite, composed of zoned plagioclase ranging from oligoclase to andesine, augite, hornblende, and biotite.)
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- + — **Pegmatite**
  (Composed chiefly of perthite with some muscovite, biotite, and quartz.)

SILURIAN? — **LOWER DEVONIAN**

- Dl / Dlg / Dl — **Littleton formation**
  (Zone b: gray phyllite and quartzite, with some mica schist and mica-quartz schist. Zone g: mica schist, mica-garnet schist, and quartzite. Zone st: mica schist, mica-garnet schist, mica-garnet-staurolite schist, and quartzite. Zone si: mica schist, mica-garnet schist, mica-sillimanite schist, and quartzite. Dlg, Gove member: chiefly silvery mica schist, some mica-staurolite schist, mica-sillimanite schist, and quartzite.)
- Sb — **Berwick formation**
  (Zone b: black, gray and green phyllite with biotite occurring both as porphyroblasts and in groundmass; quartz-mica schist; quartzite; actinolite-quartz granulite; actinolite granulite; biotite-actinolite schist; actinolite amphibolite. Zone g: similar to rocks in Zone b, but some of them contain garnet.)
- Sec / Se — **Eliot formation**
  (Zone c: dark-gray, light-gray and green phyllite; quarts-sericite schist; quartzite; calcareous phyllite. Zone b: dark-gray, light-gray, and green phyllite with biotite porphyroblasts; quartz-mica schist with biotite porphyroblasts; quartzite; actinolite-biotite granulite. Sec, Calef member: chiefly black phyllite; some green quarts-chlorite phyllite.)

METAMORPHIC ZONES

Shown below formation symbol thus: $\frac{Dl}{g}$

- c - chlorite zone
- b - biotite zone
- g - garnet zone
- st - staurolite zone
- si - sillimanite zone

CONTACTS

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- Strike of vertical beds
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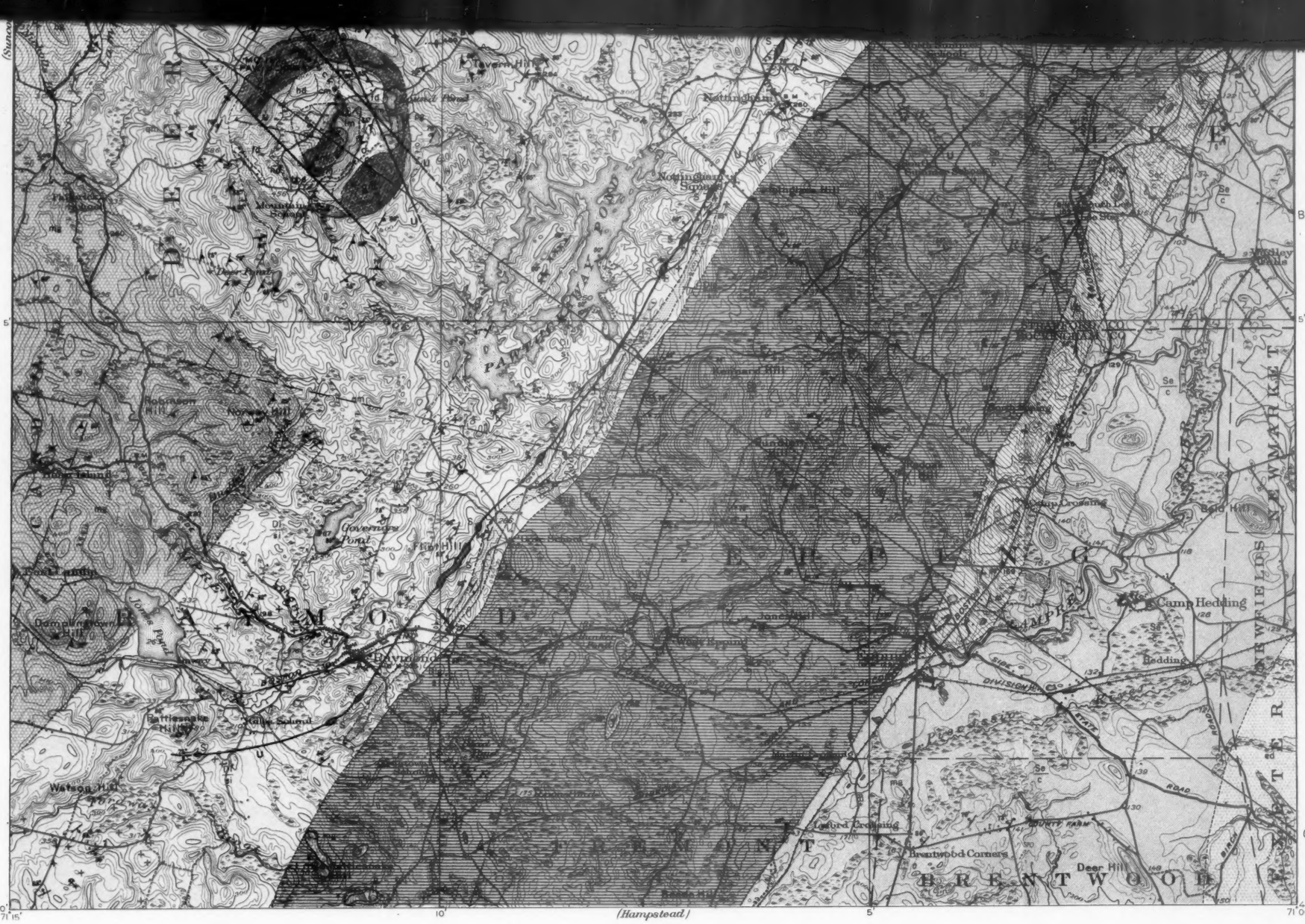


LATE DEVONIAN?

New Hampshire Magma Series

ers white. Quartz latite: bluish-gray quartz latite with phenocrysts of orthoclase and plagioclase. Keratophyre: bluish-gray keratophyre with phenocrysts of oligoclase up to 0.3 cm. long. Andesi dark-gray to blue andesite. Breccia: breccia composed of angular fragments, 1.0 to 5.5 cm. in diameter, composed of andesite, rhyolite, and schist set in a rhyolite matrix.)

g

Gabbro and monzonite

(Gabbro: coarse-grained gabbro, composed of white laths of labradorite-bytownite, with interstitial augite, olivine, and magnetite. Hypersthene gabbro: coarse-grained hypersthene gabbro, composed of labradorite, augite, magnetite, hypersthene, and biotite. Monzonite: coarse-grained hypersthene monzonite, composed of orthoclase, microperthite, oligoclase, and hypersthene.)

+

Pegmatite

(Composed chiefly of perthite with some muscovite, biotite, and quartz.)

mg

Microcline granite

(Pink to light-gray, massive to foliated, medium- grained to coarse-grained granite, composed pink microcline, quartz, muscovite, and biotite.

bg

Binary granite

(White to light-gray, medium-grained to coarse- grained, massive to faintly foliated granite an quartz monzonite, composed of microcline, orth clase, oligoclase, quartz, biotite, and muscovite

qm

Quartz monzonite

(Light-gray to dark-gray, medium-grained coarse-grained, massive to foliated quartz mon zonite and granodiorite, composed of orthoclas microcline, oligoclase, quarts, biotite, and some specimens, hornblende.)

qd

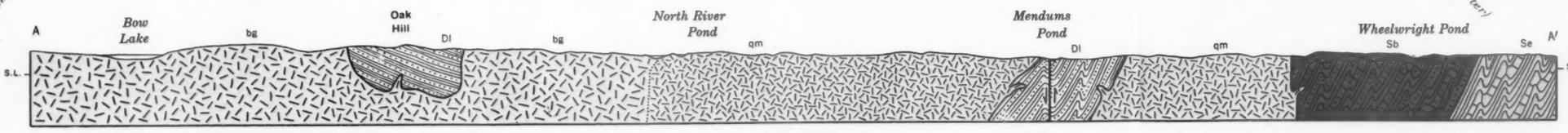
Quartz diorite

(Dark-gray, fine-grained to medium-grained, ma sive to faintly foliated rock, composed of olig clase-andesine, quartz, biotite, and hornblend

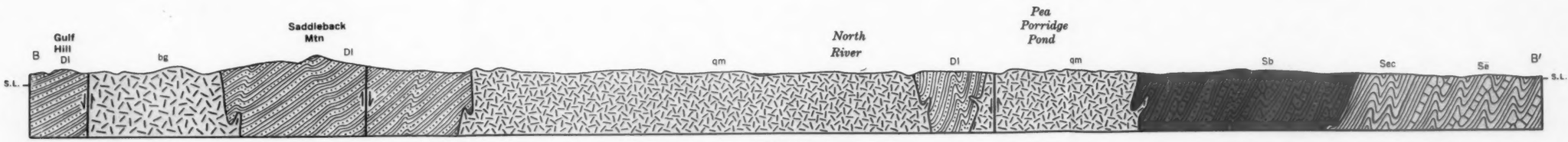
ed

Exeter diorite

(Dark-gray to black, coarse-grained, massive, cally porphyritic diorite, gabbro, and quar diorite, composed of andesine or labradorite, gite, hornblende, and a little biotite and quar



SECTION A-A'



SECTION B-B'

SECTION C-C'

GEOLOGIC MAP AND STRUCTURE SECTIONS OF THE MT. PAWTUCKAWAY QUADRANGLE, NEW HAMPSHIRE

LATE DEVONIAN?

New Hampshire Magma Series

(Gabbro: coarse-grained gabbro, composed of white laths of labradorite-bytownite, with interstitial augite, olivine, and magnetite. Hypersthene gabbro: coarse-grained hypersthene gabbro, composed of labradorite, augite, magnetite, hypersthene, and biotite. Monzonite: coarse-grained hypersthene monzonite, composed of orthoclase, microperthite, oligoclase, and hypersthene.)

Gabbro and monzonite

(Composed chiefly of perthite with some muscovite, biotite, and quartz.)

Pegmatite

(Pink to light-gray, massive to foliated, medium-grained to coarse-grained granite, composed of pink microcline, quartz, muscovite, and biotite.)

Microcline granite

(White to light-gray, medium-grained to coarse-grained, massive to faintly foliated granite and quartz monzonite, composed of microcline, orthoclase, oligoclase, quartz, biotite, and muscovite.)

Binary granite

(Light-gray to dark-gray, medium-grained to coarse-grained, massive to foliated quartz monzonite and granodiorite, composed of orthoclase, microcline, oligoclase, quartz, biotite, and in some specimens, hornblende.)

Quartz monzonite

(Dark-gray, fine-grained to medium-grained, massive to faintly foliated rock, composed of oligoclase-andesine, quarts, biotite, and hornblende.)

Quartz diorite

(Dark-gray to black, coarse-grained, massive, locally porphyritic diorite, gabbro, and quartz diorite, composed of andesine or labradorite, augite, hornblende, and a little biotite and quarts.)

Exeter diorite

Strike and dip of bedding

Strike of vertical beds

Horizontal bedding

Strike and dip of foliation of igneous and metamorphic rocks

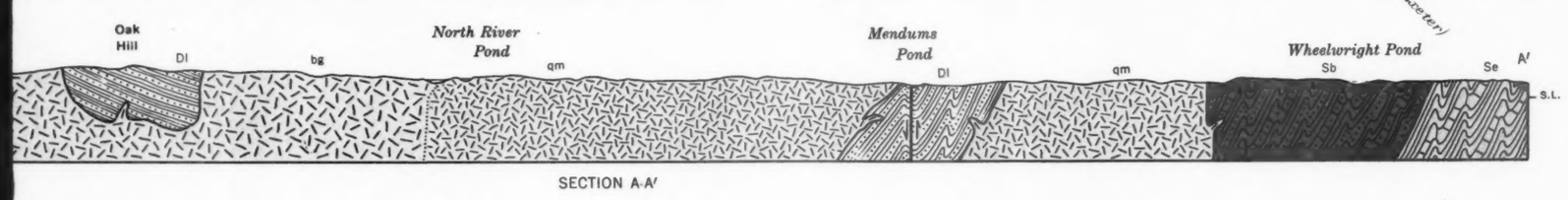
Strike of vertical foliation of igneous and metamorphic rocks

FAULT

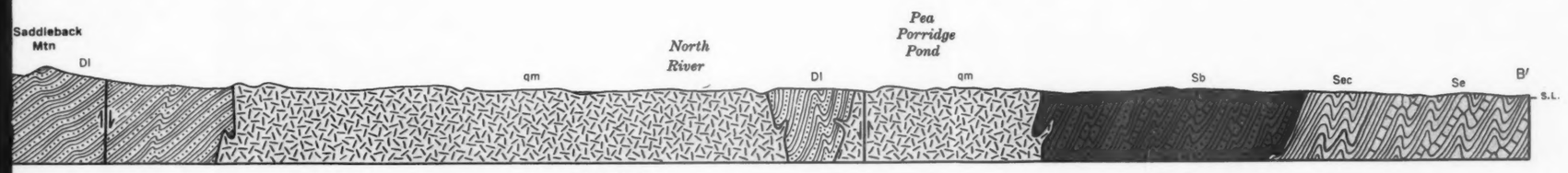
"U" indicates upthrown side

Silicified zone

Quarries, mostly abandoned



SECTION A-A'



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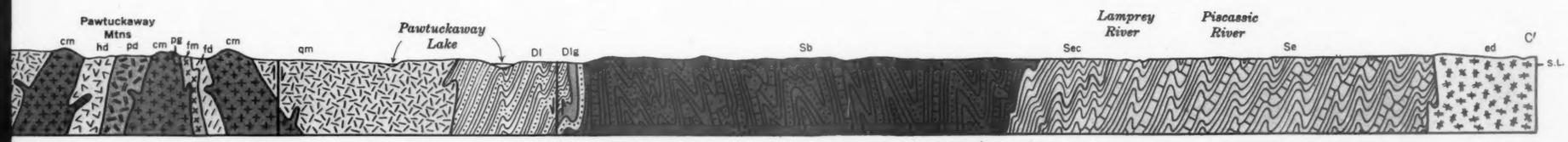
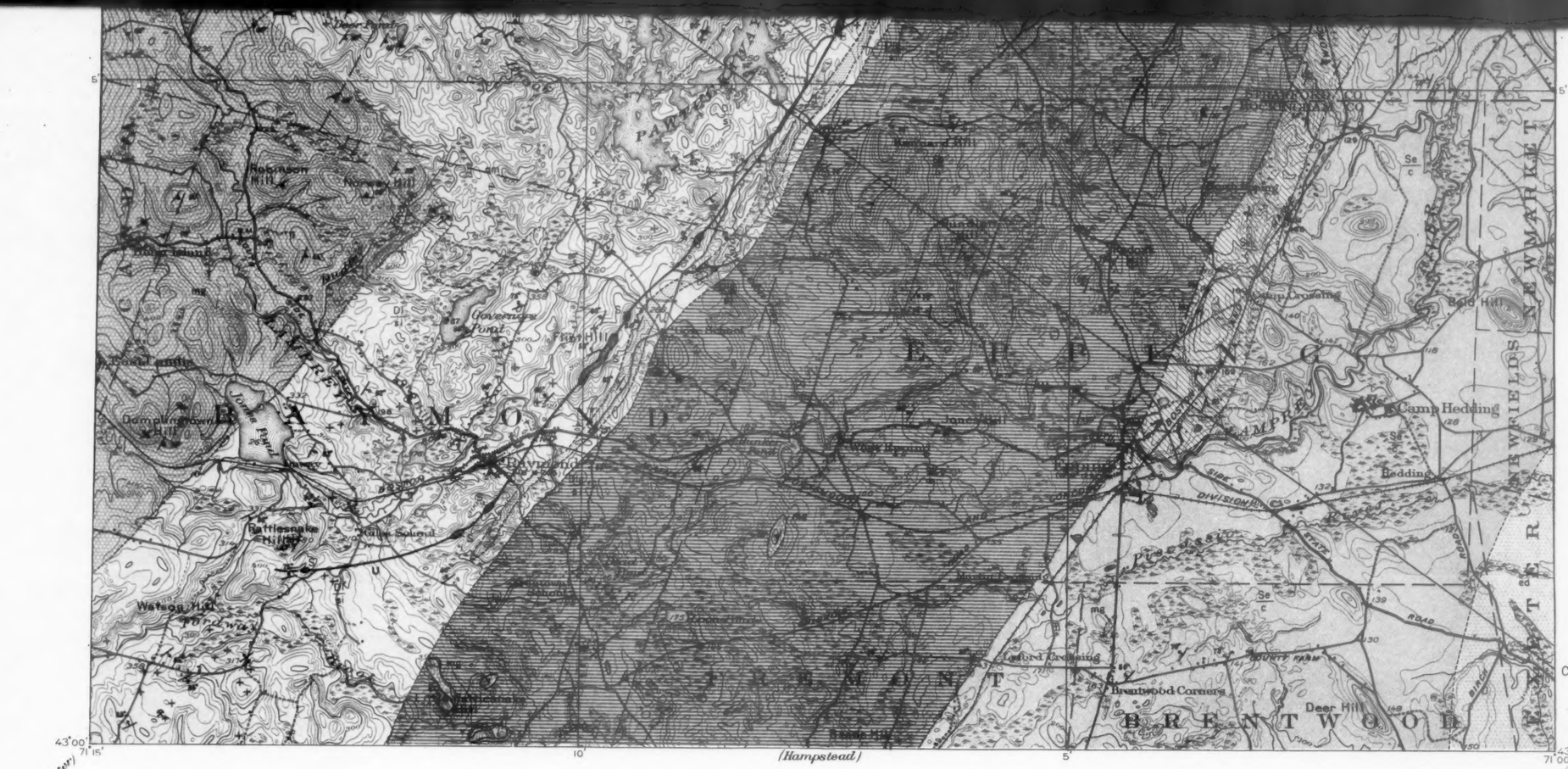


SECTION C-C'

STRUCTURE SECTIONS OF THE MT. PAWTUCKAWAY QUADRANGLE, NEW HAMPSHIRE

Scale 1/62500

1 2 3 4 Miles



SECTION A-A′

SECTION B-B′

SECTION C-C′

GEOLOGIC MAP AND STRUCTURE SECTIONS OF THE MT. PAWTUCKAWAY QUADRANGLE, NEW HAMPSHIRE

Topographic base by the U. S. Geological Survey in cooperation with the War Department.

Scale 1/62500

1 0 1 2 3 4 Miles

5000 0 5000 10000 15000 20000 Feet

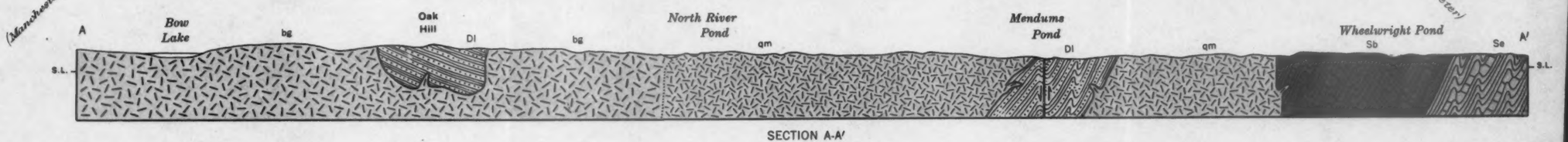
1 0 1 2 3 4 5 Kilometers

Contour interval 20 feet

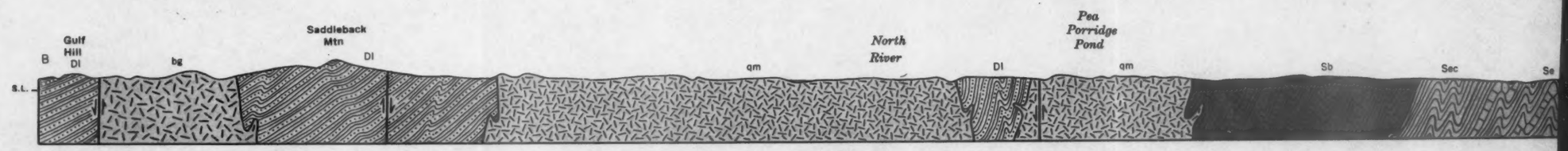
Datum is mean sea level

APPROXIMATE MEAN DECLINATION 1917

Geology by Jacob Freedman assisted by T. Botinelly. Directed by Marland P. Billings. Geology surveyed in 1940, 1941, and 1946. Published in 1950.



SECTION A-A'



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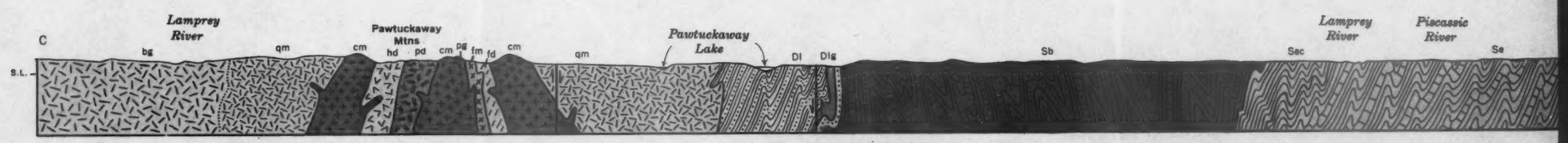
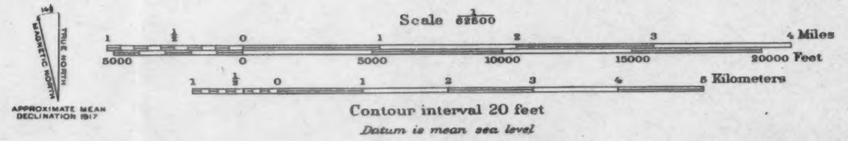


SECTION C-C'

GEOLOGIC MAP AND STRUCTURE SECTIONS OF THE MT. PAWTUCKAWAY QUADRANGLE, NEW HAMPSHIRE

Topographic base by the U. S. Geological Survey in cooperation with the War Department.


MAGNETIC NORTH
TRUE NORTH
APPROXIMATE MEAN DECLINATION 19.7°
Scale 62500
1 0 1 2 3 4 Miles
5000 0 5000 10000 15000 20000 Feet
1 0 1 2 3 4 Kilometers

Contour interval 20 feet

Datum is mean sea level

Geology by Jacob Freedman assisted by T. Botinelly. Directed by Marland P. Billings. Geology surveyed in 1940, 1941, and 1946. Published in 1950.

LATE DEVONIAN?

New Hampshire Magma Series

Pegmatite

(Composed chiefly of perthite with some muscovite, biotite, and quartz.)

mg

Microcline granite

(Pink to light-gray, massive to foliated, medium-grained to coarse-grained granite, composed of pink microcline, quartz, muscovite, and biotite.)

bg

Binary granite

(White to light-gray, medium-grained to coarse-grained, massive to faintly foliated granite and quartz monzonite, composed of microcline, orthoclase, oligoclase, quartz, biotite, and muscovite.)

qm

Quartz monzonite

(Light-gray to dark-gray, medium-grained to coarse-grained, massive to foliated quartz monzonite and granodiorite, composed of orthoclase, microcline, oligoclase, quartz, biotite, and in some specimens, hornblende.)

qd

Quartz diorite

(Dark-gray, fine-grained to medium-grained, massive to faintly foliated rock, composed of oligoclase-andesine, quartz, biotite, and hornblende.)

ed

Exeter diorite

(Dark-gray to black, coarse-grained, massive, locally porphyritic diorite, gabbro, and quartz diorite, composed of andesine or labradorite, augite, hornblende, and a little biotite and quartz.)

Strike and dip of foliation of igneous and metamorphic rocks

Strike of vertical foliation of igneous and metamorphic rocks

FAULT

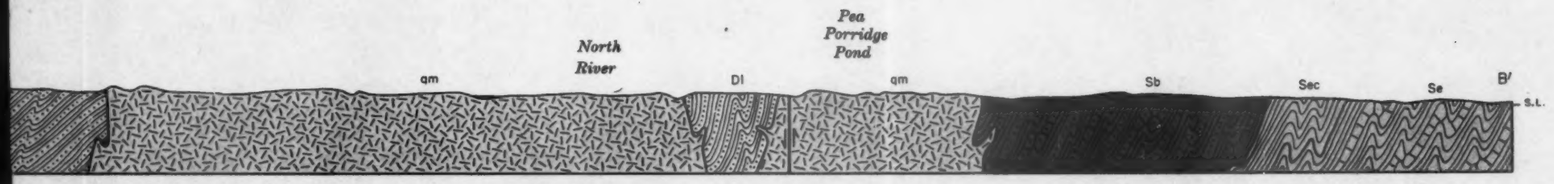
"U" indicates upthrown side

S

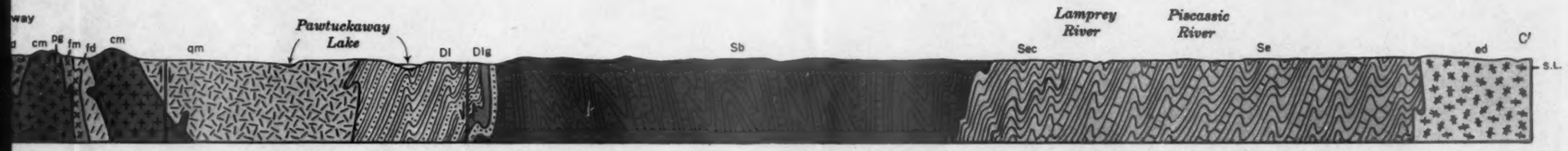
Silicified zone

Quarries, mostly abandoned

SECTION A-A'



SECTION B-B'



SECTION C-C'

RE SECTIONS OF THE MT. PAWTUCKAWAY QUADRANGLE, NEW HAMPSHIRE

Scale 1/62500

1 Miles

5000 Feet

5 Kilometers

Contour interval 20 feet

Datum is mean sea level

Geology by Jacob Freedman assisted by T. Botinelly. Directed by Marland P. Billings. Geology surveyed in 1940, 1941, and 1946. Published in 1950.

LATE DEVONIAN?

New Hampshire Magma Series

Pegmatite

(Composed chiefly of perthite with some muscovite, biotite, and quartz.)

mg — **Microcline granite**

(Pink to light-gray, massive to foliated, medium-grained to coarse-grained granite, composed of pink microcline, quartz, muscovite, and biotite.)

bg — **Binary granite**

(White to light-gray, medium-grained to coarse-grained, massive to faintly foliated granite and quartz monzonite, composed of microcline, orthoclase, oligoclase, quartz, biotite, and muscovite.)

qm — **Quartz monzonite**

(Light-gray to dark-gray, medium-grained to coarse-grained, massive to foliated quartz monzonite and granodiorite, composed of orthoclase, microcline, oligoclase, quartz, biotite, and in some specimens, hornblende.)

qd — **Quartz diorite**

(Dark-gray, fine-grained to medium-grained, massive to faintly foliated rock, composed of oligoclase-andesine, quartz, biotite, and hornblende.)

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(Dark-gray to black, coarse-grained, massive, locally porphyritic diorite, gabbro, and quartz diorite, composed of andesine or labradorite, augite, hornblende, and a little biotite and quartz.)

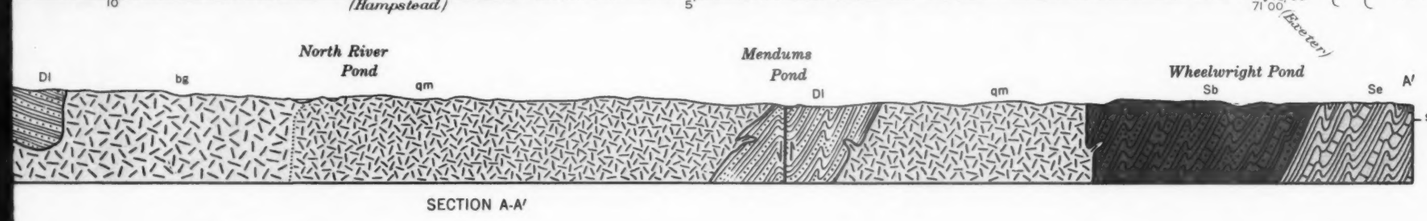
Strike and dip of foliation of igneous and metamorphic rocks

Strike of vertical foliation of igneous and metamorphic rocks

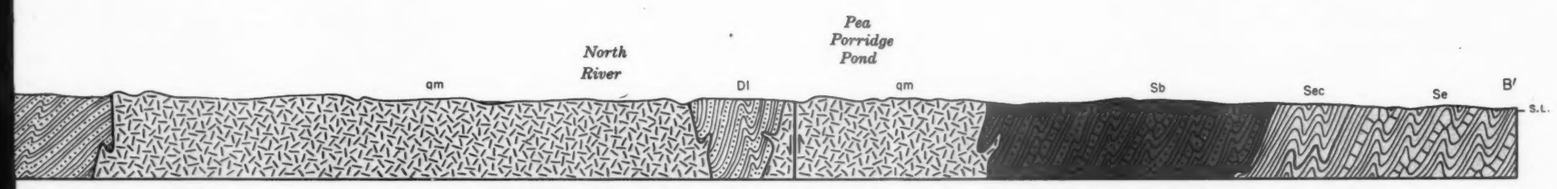
FAULT
"U" indicates upthrown side

S — Silicified zone

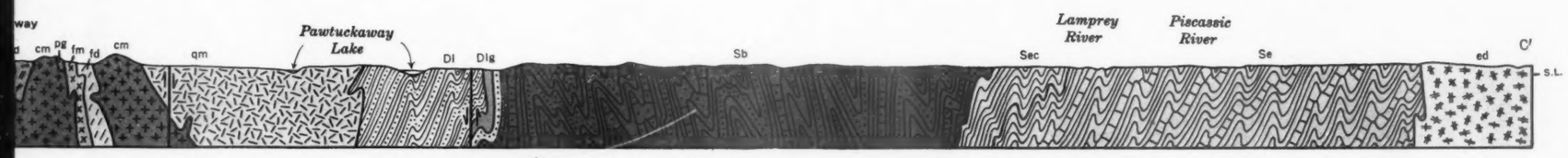
Quarries, mostly abandoned



SECTION A-A'



SECTION B-B'



SECTION C-C'

IRE SECTIONS OF THE MT. PAWTUCKAWAY QUADRANGLE, NEW HAMPSHIRE

Scale $\frac{1}{62500}$

1 ½ 0 1 2 3 4 Miles

5000 0 5000 10000 15000 20000 Feet

1 ½ 0 1 2 3 4 5 Kilometers

Contour interval 20 feet

Datum is mean sea level

E MEAN
N 1917

Geology by Jacob Freedman assisted by T. Botinelly. Directed by Marland P. Billings. Geology surveyed in 1940, 1941, and 1946. Published in 1950.

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of potash feldspar. There Billings (1941) says the scarcity is because "... not enough potash was introduced to change the muscovite to potash feldspar."

Orthoclase comprises 10 to 35 per cent of two specimens of lime-silicate rock, and microcline constitutes 30 per cent of one specimen. These probably formed from recrystallized detrital feldspar, or from the reaction of sericite and dolomite, producing microcline, actinolite, and anorthite.

Trachyte is the only dike rock containing potash feldspar. Potash feldspar is abundant in the New Hampshire magma series and the White Mountain magma series.

Structure

General Statement

The Mt. Pawtuckaway quadrangle is on the southeast limb of a major syncline, the axis of which lies to the northwest of the quadrangle. Minor folds are superimposed on the major structure. Plutonic rocks of the New Hampshire magma series and the White Mountain magma series transect structures.

A well-developed schistosity consistently parallels the bedding in the argillaceous rocks. Lineation formed by crinkles and aligned mica flakes on the foliation planes are essentially parallel to the axes of the minor folds. Fracture cleavage is developed locally and is essentially parallel to the axial planes of the minor folds.

A series of silicified zones, some of which are aligned, are presumably faults. Three demonstrable faults offset the rocks in the Pawtuckaway Mountains, and others occur in the Saddleback Mountain and in the Gulf Hill area.

The regional trend of the rocks in the Mt. Pawtuckaway quadrangle is N.30°E. (Pl. 1). Bedding is well exposed in most outcrops. In the southeastern half of the quadrangle the strike of the bedding ranges from N.15°E. to N.70°E. The beds dip steeply northwest and southeast and the angles of dip range from vertical to 50°. Drag folds indicate locally that the beds are overturned toward the northwest. In the Saddleback Mountain area, the beds strike north and northwest and dip from 20° to 55° westward. Bedding of the Littleton formation in the Gulf Hill area generally strikes north-south, but the strike ranges from N.50°W. to N.40°E.; the dip of the beds ranges from 25° westward to vertical.

Major Fold

General statement.—Southeast of the large area of plutonic rocks that occupies much of the quadrangle, data on drag folds and lineation suggest that the folds plunge northeast, but a broad structural terrace in the Oak Hill-Saddleback Mountain area has an opposite plunge to the southwest. Evidence for the direction of plunge of the major syncline is conflicting, but drag folds and rare fracture cleavage indicate that the top of the beds is to the northwest in six cases for every one to the southeast. Evidence for this interpretation is discussed below.

Stratigraphic position of the Berwick formation.—It has been indicated that the metamorphosed, sedimentary rocks of the Mt. Pawtuckaway quadrangle (Pl. 1) form belts that trend north-northeast and that they have been assigned to the Eliot, Berwick, and Littleton formations successively to the northwest. A general northwesterly dip is rather clear in spite of local differences. Each formation is distinctive lithologically, and there is no evidence that the formations are repeated.

The relationship of the Berwick formation to the Eliot and Littleton formations is of critical importance because, although Katz's map (1917) portrays the distribution of lithologic types accurately, his interpretation of the age of the Berwick formation is doubtful. He referred the Berwick to "pre-Algonkian" because "its greater deformation and more complex structure, its notably greater metamorphism, and the presence in it of deformed intrusives not known to invade other formations show that the Berwick gneiss is older than the sedimentary formations which adjoin it."

The "greater deformation" of the Berwick formation is difficult to find. The beds in the Berwick formation are no more folded or faulted than those in the Eliot or Littleton formation. The argillaceous layers interbedded with quartzite are minutely folded, but no more so than the phyllite in the Eliot formation. The "more complex structure" was not evident inasmuch

as all of the formations responded in a similar manner to deformation, and all are on the same limb of one fold. "Its notably greater metamorphism" was found to be the normal increase in metamorphism from the chlorite zone in the Eliot formation to the biotite zone in the Berwick formation. Actually most of the Littleton formation is of higher metamorphic grade than the Berwick formation. "Deformed intrusives not known to invade other formations" were impossible to find in the Mt. Pawtuckaway quadrangle. Katz refers to these as follows: "In the northwestern part of the region (of his study) there are bodies of intrusive gneissic granite and pegmatite, which are restricted to the area of the Berwick." His map indicates that the intrusive rock is "Biddeford granite and other granites, granite gneiss, and pegmatite." The description is similar to that of rocks of the New Hampshire magma series in the Mt. Pawtuckaway quadrangle. Locally, near contacts, the granitic rocks are more gneissose, a feature related to the mode of intrusion. Katz may have interpreted these as deformed intrusives.

The writer's studies show that the contact between the Berwick formation and the Eliot formation is gradational. The bedding and schistosity are parallel in both formations and there is no sign of a fault or unconformity. It is therefore concluded that the Berwick formation is conformable with the Eliot formation. The same conclusion has been reached independently in the Dover quadrangle to the east (T. R. Meyers, oral communication).

The contact of the Berwick formation with the Littleton formation is marked by numerous outcrops of actinolite granulite and biotite-actinolite schist at the top of the Berwick formation and by mica schist, mica-staurolite schist and mica-sillimanite schist at the base of the Littleton formation. Again the structure in adjacent formations is parallel and there is no evidence of a fault or unconformity.

It is apparent, therefore, that the strata in the Mt. Pawtuckaway quadrangle belong to the limb of a single fold. Determination of the direction in which beds become younger has been inhibited by the absence of criteria for determining tops of beds, such as fossils, cross-bedding or graded bedding; fracture cleavage is rare. Two other lines of evidence, however, indicate that the tops of beds are to the northwest and that the strata are in the southeast limb of a major syncline.

Minor folds.—The use of drag folds to determine the position of rocks of isolated exposures relative to major fold axes is well known (Billings, 1942, p. 76–81). Data were recorded on minor folds in 70 different localities in the Mt. Pawtuckaway quadrangle. At 34 of these localities, the exposure was too poor to indicate direction of beds. Of the other 36, however, 30 indicate that the tops of beds are to the northwest, and 6 indicate that the tops are to the southeast. In the area southeast of the central pluton, 85 per cent of the folds plunge 15° to 80° northeast; 10 per cent plunge 4° to 45° southwest; and 5 per cent are horizontal.

Saddleback Mountain structural terrace.—The strata on Saddleback Mountain and Oak Hill strike northwest at right angles to the regional trend and dip southwest, apparently at the nose of a plunging fold. This appears to be a large drag fold (Pl. 1, structure sections) indicating tops to the northwest, but the direction of plunge conflicts with that of folds in the southeastern part of the quadrangle.

New Hampshire Magma Series

General relations.—Exposures of the Exeter diorite are too few in the Mt. Pawtuckaway quadrangle to provide any evidence of the structure.

A body of quartz diorite extends northeast from near West Barrington. Near the contact with the Littleton formation at the north end of Mt. Misery, the foliation dips 30° N.W. although, in general, the quartz diorite in this body is massive. The contact presumably has a similar dip, therefore the body is apparently a thick sheet concordant with the underlying schists. The schists exposed in the extreme north-central part of the quadrangle, east of Strafford Post Office, presumably lie above the sill of quartz diorite.

There is a second body of quartz diorite in the extreme northeast corner of the quadrangle. In the vicinity of Poor Farm School, the foliation of the quartz diorite dips steeply northwest toward the schist. Assuming that the

contact has a similar attitude, this body of quartz diorite is a sheet-like mass underlying the schist. Around Berrys Hill, however, the foliation dips east. The sheet of quartz diorite thus appears to occupy the core of an anticline that plunges north.

In the northeast corner of the quadrangle an anticlinal body of quartz diorite is overlain concordantly by the schists that extend northeast from West Barrington. These schists are in turn overlain by a second sill of quartz diorite that extends north of Nippo Hill.

The northeastern part of the quartz monzonite is generally massive. From Kenison Pond in Nottingham southeastward, the foliation suggests a partial dome (Pl. 1) plunging 30° to 50° northeast. Foliation is common in the quartz monzonite around the Pawtuckaway Mountains. "The foliation of the gneiss does not conform to the outline of the intrusive complex (Roy and Freedman, 1944)." The strike of the foliation is generally northeast, parallel to the regional trend, and the dips range from northwest to southeast. In a number of readings however the strike ranges from east-west to northwest.

The binary granite is dominantly massive. Foliation planes in the vicinity of Northwood Narrows and Deerfield Center strike generally east-west and northwest, and dip from vertical to 20° north and south. A few scattered readings on foliation planes have no consistent orientation.

Foliation is uncommon in the microcline granite. It is produced by broadly waving and irregular concentrations of biotite. In general, the foliation strikes northeast and dips from vertical to 60° north and east.

Mechanics of intrusion.—Data on the Exeter diorite are insufficient to discuss its mode of emplacement. Its composition suggests that it is related to an early stage in the intrusion of the New Hampshire magma series.

The bodies of quartz diorite in the northeastern part of the quadrangle appear to be concordant sheets which were injected as large sills, in a manner similar to the Mt. Clough pluton of western New Hampshire (Billings, 1945, p. 57–62).

The method of formation of the large body of plutonic rock in the central and northwestern part of the quadrangle is an intriguing problem. Methods by which quartz monzonite, binary granite, and microcline granite of the New Hampshire magma series could have been emplaced include the general hypotheses of forceful injection, granitization, permissive intrusion, and piece-meal stoping.

If the magma had been forcefully injected, one would expect bedding in the country rock wrapping concordantly around the igneous bodies, and foliation planes in the igneous rock abundant at and parallel to the contacts. The southeastern contact of the quartz monzonite and microcline granite, that trends northeast from the southwest corner of the map to the northeast corner, is clearly not concordant. Various tongues and lobes cut discordantly across the metamorphosed sedimentary rocks. This is apparent southwest of Pawtuckaway Pond. A more striking example of discordant relations is shown in the eastern part of Nottingham township, where a great lobe bites deeply into the metamorphic rocks. The contacts elsewhere, notably north of West Barrington and north of Saddleback Mountain, show similar discordant relations. Forceful injection is not plausible for these.

Granitization has been reported in the Chelmsford - Westford area in Massachusetts (Currier, 1947) which is on the regional strike approximately 30 miles southwest of the Mt. Pawtuckaway quadrangle. A number of peculiarities in the granitic body in the Mt. Pawtuckaway quadrangle make it necessary to consider granitization as a mode of emplacement. Read (1944, p. 46) has stated that, "Granitization means the process by which solid rocks are converted to rocks of granitic character without passing through a magmatic stage."

A glance at the map of the Mt. Pawtuckaway quadrangle (Pl. 1) reveals a number of isolated bodies of metamorphic rocks of varying size in the body of plutonic rock. Most of these have been modified, presumably by magma or by emanations from magma from which the plutonic rocks formed.

The metamorphic rocks at the northeast end of the Saddleback Mountain area are tourmalinized to a great extent. Sprays of radiating crystals occur along bedding planes, larger

individual crystals are scattered in the schist, and locally the rock is almost completely replaced by fine-grained tourmaline. On the road southwest of Gove School in the northeastern part of Raymond township, veins 1–2 inches thick are composed of a compact aggregate of small tourmaline needles. These are believed to have been formed by aqueous solutions rich in boron given off by the magma in the late stages of solidification.

In the smaller isolated islands of metamorphic rock, igneous-looking material occurs along the foliation planes varying from a fraction of an inch to a few inches thick. These features produce a rock that answers the description of migmatites or mixed rocks (Sederholm, 1926). They consist of irregularly folded aggregates of igneous-looking and metamorphic-looking minerals occurring in numerous outcrops between Saddleback Mountain and Oak and Bennett Hills in Northwood township.

Much of the quartz monzonite and microcline granite plutons contain very little biotite. In general the biotite occurs scattered through the rock with a faint orientation. Locally, however, schlieren of thickly matted, coarse-grained biotite varying in length and thickness occur in the granitic rock. These may have been segregated in the magma and dragged out by magmatic flow. On the other hand, they occur in rocks that in some outcrops show no other sign of flow and suggest layers of country rock, engulfed in the magma and recrystallized to coarser-grained biotite.

Hornblende is not commonly found in the granitic rocks less mafic than granodiorite, but in large outcrops of microcline granite at Robinson Hill in Raymond and on the South Side State Road west of the Candia-Raymond township line, hornblende occurs in scattered grains averaging 0.5 inch in diameter. The hornblende is almost always associated with pink microcline which occurs in large scattered grains and in lenses as much as 6 inches thick.

The rocks of the large outcrops of microcline granite are quite variable in composition and structure. Biotite occurs with hornblende in local areas scattered through the rock. It also occurs without hornblende in numerous foliation planes spaced about 0.1 inch apart. Areas with biotite foliation are separated from rock of different mineral composition and structure by sharp boundary lines. Parallel, thin, discontinuous layers of folded biotite separated by granite form another occurrence. The zones in which the biotite foliation planes are closely spaced seem to be inclusions of country rock in the granite which form injection gneisses. These inclusions are rotated locally out of parallelism with foliation planes in granite. Where the biotite foliation planes are more widely spaced and separated by thicker layers of granite, magma has presumably been injected. Outcrops in which biotite and hornblende are abundant are believed to be inclusions of the country rock that have been reworked by magma.

Feldspar is abundant in only a portion of the metamorphic rocks. Most such occurrences are in the Littleton formation within a mile of the igneous contact. They contain plagioclase feldspar ranging up to 40 per cent of the rock and as much as 65 per cent perthite. Most of the feldspar, however, is fine-grained and does not form porphyroblasts as might be expected if the rocks had been granitized. Moreover, plagioclase in the igneous rocks is zoned, indicating a magmatic origin (Gilluly, 1948). Thus, although inclusions of schist have in many localities been converted to igneous-looking rocks—that is, granitized—these are local phenomena to be expected for inclusions caught in a magma; they are not part of an overall granitization.

Some of the New Hampshire magma series in other localities (Billings, 1937) have been emplaced in part by permissive intrusion. The close structural parallelism between country rock and plutonic rock typical of this mode of emplacement is found locally at Pawtuckaway Pond, in Barrington township and at isolated patches of Littleton formation in quartz monzonite and binary granite. These rocks were apparently emplaced to some extent by permissive intrusion synchronous with the orogeny.

A consideration of piecemeal stoping follows. Observed contacts between plutonic rocks and metamorphic rocks are invariably sharp with interfingering of the two rocks in a contact zone. Reference has already been made to the scalloped character of the southeastern contact

of the quartz monzonite and the contact northeast of Saddleback Mountain.

One contact was observed on the north side of the road 0.75 mile northwest of Nottingham Square, but interfingering of plutonic and metamorphic rock is best exposed on the southwest shore of Pawtuckaway Pond, west of Langley Pond in Nottingham township, and at Mendums Pond in Barrington township. The contact of the main belt of the metamorphic rocks and the granitic rocks runs through 5 lakes, whose basins were eroded along the weakened contact zone.

Northwest of Nottingham Square, the plutonic rock is intruded parallel to the schistosity with a knife-edge sharp contact. No outcrops were observed to the northeast to indicate whether the sharp contact persists or whether an interfingering of granitic and metamorphic rock occurred.

At Pawtuckaway Pond and Mendums Pond, and west of Langley Pond, large scale interfingering marks a broad contact zone. Locally, the granite cross-cuts the country rock, but, for the most part, bands from a fraction of a foot to 100 feet thick are interfingered. The contacts between layers are invariably sharp. Numerous inclusions of the country rock in the granitic rock were observed. Some were parallel to the regional trend, but many had irregular shapes and no conformable orientation. Near the contacts the inclusions were affected largely by an increase in grain size and the addition of potash, forming coarse-grained muscovite. In the interior of the plutonic mass, the changes made in the inclusions ranged from the injection of a little granitic material along a few foliation planes to complete granitization.

Where the country rock is interfingered with the granitic rock, the magma came up along lines of easiest access, the bedding and foliation planes in the metamorphic rocks. Large sections of the country rock were cut off and engulfed. Apophyses of granitic rock also cross-cut the strata in numerous places. These are generally much smaller than the interfingered granitic rock, but locally they constitute areas a few hundred feet across.

The large embayment of plutonic rock in the east-central ninth of the quadrangle cross-cuts the Berwick formation for about 3 miles. Although the contacts of the embayment were not observed there can be little doubt that they are cross-cutting.

There are no gradations in composition between plutonic and metamorphic rock in the Mt. Pawtuckaway quadrangle. Nor are there any porphyroblasts of feldspar in the metamorphic rocks to indicate incipient replacement. The relationship of the plutonic rock to the metamorphic rock is definitely intrusive and in numerous places cross-cutting. The evidence is strongly in favor of piecemeal stoping, and locally permissive intrusion as the mechanisms of emplacement of the plutonic rocks.

Silicified Zones and Faults

Silicified zones.—The silicified zones are similar to those found along faults in other parts of New Hampshire. The writer has observed three lines of silicified zones, designated the Flint Hill, the Nottingham, and the Calef Road. The Saddleback Mountain and the Gulf faults are also described.

The Flint Hill silicified zones may be traced for 18 miles from Rattlesnake Hill in Raymond township to Huckleberry Hill in Barrington township. From Rattlesnake Hill to Flint Hill the line of silicified zones swings from an east-west strike to N.50°E. North of Flint Hill the strike is N.30°E. The Flint Hill silicified zones, which are the longest series in the quadrangle, are marked by 12 exposures of silicified rock. These are either oval areas of milky quartz, country rock veined or almost completely replaced by quartz, or country rock silicified to a chert. The maximum length of an exposed silicified zone is approximately 3,250 feet, but they average about 500 feet long and range from 10 to 100 feet wide. A few silicified zones are apparently made up of two bodies which are en echelon.

This line of silicified zones is believed to have formed along a fault, which is believed to be essentially vertical (Sections AA′, BB′, and CC′, Pl. 1) since topography appears to exert no influence on its trace. It is difficult to determine the direction and amount of the net slip, but one line of evidence indicates that the northwest side of the fault is downthrown. The Gove member of the Littleton formation

crops out southeast of the fault (Pl. 1). If the northwest side of the fault were upthrown, the Gove member would almost certainly be repeated northwest of the fault. It appears, therefore, that the block southeast of the fault is upthrown, as shown in Section BB′, Pl. 1. The contact, as shown on the map (Pl. 1), between the quartz monzonite and the Littleton formation is offset along the fault northeast and southwest of Mendums Pond and near Langley Pond to agree with this conclusion, although the contact at these places is not exposed.

The Nottingham line of silicified zones extends from the southeastern corner of the central-ninth of the quadrangle to Marston Pond. This zone is marked by three silicified areas, two composed of silicified country rock and one a small zone of milky quartz. Another silicified zone is represented by one outcrop 1 mile northwest of South Lee.

The Calef Road silicified zone lies in the southeast corner of the northeast ninth of the quadrangle. In a belt approximately 100 feet long and trending N.10°E. a few aligned outcrops of country rock are brecciated, silicified and veined with milky quartz.

The southeast side of the faults indicated by these silicified zones is indicated as upthrown (Pl. 1) but direct field evidence is lacking and the displacement is unknown.

Saddleback Mountain fault.—The Saddleback Mountain fault, located along the southeast edge of Saddleback Mountain strikes N.55°E. The fault was not observed in the field but is indicated by the structure, lithological change, and differences in metamorphism. Rocks of the Littleton formation occur on both sides of the fault.

The structure southeast of the fault is parallel to the regional trend with beds striking northeast and dipping northwest. Northwest of the fault the beds dominantly strike northwest at right angles to the regional trend and dip southwest. Because the rocks on both sides of the fault were not observed in contact in the field, its attitude can be inferred only from topographic expression, lack of which here suggests a vertical or steep northwest dip.

Metamorphism normally decreases northwestward in the northwest part of the quadrangle. Rocks southeast of the fault are dominantly thin-bedded, quartz-biotite schist. Northwest of the fault they are chiefly poorly-bedded, quartz-staurolite schist. Thus a local reversal in the pattern of the metamorphic zones is apparently indicated. This can best be explained by having the northwest block upthrown. The throw is unknown but the difference in the intensity of metamorphism suggests that it is considerable.

The Gulf fault.—The Gulf fault extends at least 1 mile, and possibly farther, southeast from Suncook Pond in Northwood township, striking N.20°W. It separates the Gulf Hill area of the Littleton formation from the binary granite of the New Hampshire magma series. It is expressed topographically by the steep western scarp of The Gulf, and is also indicated by breccia in the scarp. The Littleton formation in the Gulf Hill area is composed of biotite-bearing quartzites and phyllites, rocks of a lower grade of metamorphism than the nearest metamorphic rocks east of The Gulf fault. This relationship suggests that the block west of the fault is downthrown; the amount of throw is not known.

Pawtuckaway Mountains Complex

General statement.—The Pawtuckaway Mountains are composed dominantly of a core of dioritic rocks surrounded by a ring-dike of coarse-grained monzonite (Roy and Freedman, 1944). The surrounding igneous rock is quartz monzonite of the New Hampshire magma series. The arcuate body of monzonite was emplaced by the mechanism of cauldron subsidence and ring-dike injection comparable to the mechanism described by Anderson (1936).

Transverse faults.—Faults cut the Pawtuckaway Mountains into a number of segments. Two scarps, each of which has a relief of approximately 200 feet, at the north end of Mt. Pawtuckaway (Pls. 1, 2), were originally thought to have been produced by joints (Roy and Freedman, 1944, p. 914). After aerial photographs became available it was apparent that both of these scarps are in line with faults which cross the Pawtuckaway Mountains area and intersect on a steep slope west of the south end of Round Pond. The faults cross varied

AERIAL PHOTOGRAPH OF THE PAWTUCKAWAY MOUNTAINS

topography without any deviation, indicating that they are vertical. The fact that the line of the fault that strikes N.44°W. is continuous whereas the one that strikes N.25°W. seems to fade at the point of intersection suggests that the latter formed first. There is no apparent offset of either fault at the intersection. This is interpreted as indicating that the movement on the faults was vertical. A third fault, not so obvious, that strikes N.41°W., apparently cuts across the southwest extremities of the coarse-grained monzonite bodies.

It was originally assumed that the peripheral contacts of both the diorite and the coarse-grained monzonite dip outward at the same angle everywhere. Actual determinations could be made at only two points, at the outer contacts of the coarse-grained monzonite (Roy and Freedman, 1944), and they show outward dips of 55° and 80°. Irregularities in the topography and in the outside contact of the Mt. Pawtuckaway complex were used to determine the offset of the contacts along the faults. If 55° is used as the dip of the contact, the net slip of the faults is approximately 300 feet. If 80° is used, the net slip is approximately 1,000 feet. The block northeast of the fault that strikes N.25°W. and the block southwest of the fault that strikes N.44°W. were down-dropped. The net slip on the third fault is apparently about the same as on the other two; the southwest block was downthrown.

Roy and Freedman (1944, p. 913) proposed that "The outer arc of monzonite was later injected between the diorite and the gneiss for three-quarters of its circumference . . . " It now seems more likely that a complete ring of quartz monzonite was intruded part way into the diorite. The south and west portion was downfaulted below the top of the quartz monzonite so that only the overlying diorite is exposed (Fig. 7).

The sequence of events following the injection and crystallization of the Mt. Pawtuckaway complex is as follows:

(1) Vertical faulting along a line striking N.25°W. The northeast block is downthrown.

(2) Vertical faulting along a line striking N.44°W., intersecting the first fault. The southwest block is downthrown.

(3) Faulting across the southwest ends of the coarse-grained monzonite bodies. The southwest block is downthrown.

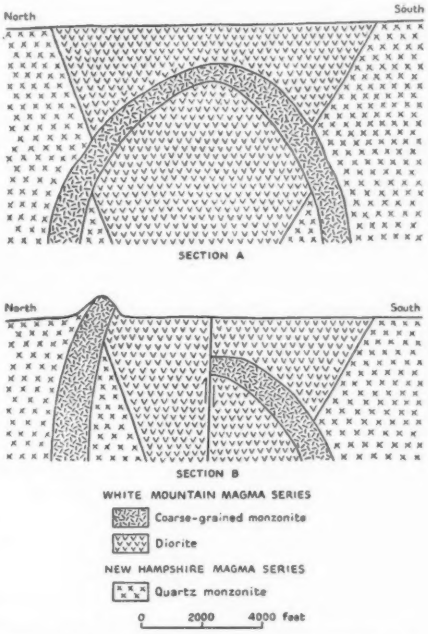

FIGURE 7.—DIAGRAMMATIC CROSS SECTION OF THE PAWTUCKAWAY MOUNTAINS

A. Intrusion of the coarse-grained monzonite by cauldron-subsidence. B. Structure after faulting and erosion.

Mechanics of intrusion.—Roy and Freedman (1944) favored the view that the diorite body was the result of "the subsidence into a magma chamber of an essentially circular block bounded by fractures which dip steeply outward. The magma welled up around and over this block producing the diorite body with local gabbroic facies." Billings (1945, p. 56) states:

"The gabbro of the White Mountain magma series, found at . . . the Pawtuckaway Mountains, was probably forcefully injected pushing aside the older rocks. At . . . the Pawtuckaway Mountains, the inward dip of the banding of the gabbro suggests that these intrusions occupy funnels that narrow downward. Balk (1937, p. 92–95) has discussed this characteristic of gabbro massifs elsewhere. Moreover, it is interesting to note that the average specific gravity of the crystalline rocks older than the White Mountain magma series would be of the

order of 2.55, whereas the specific gravity of gabbro magma would be of the order of 2.70 (Daly, 1933, p. 276). Thus, during the gabbro stage, stoping of any kind would be impossible.

"When the magma had evolved to monzodiorite, however, it would have a lower specific gravity than the overlying rocks, and stoping would be possible."

On theoretical grounds, and by comparison with the gabbro stage of the White Mountain magma series at Mt. Tripyramid, Mt. Ascutney, and the Belknap Mountains (Billings, 1945, p. 56) the diorite body in the Pawtuckaway Mountains was probably forcefully injected. Roy and Freedman (1944, p. 916) admit that "there remains the possibility that it is a stock which forced aside the older rock to reach its present level." They discard this idea because the elliptical plan of the diorite body is elongate across rather than parallel to the regional trend; and the foliation in the older gneiss does not wrap around the intrusion. The diorite body is more nearly circular than elliptical, and has no longer axis. The foliation of the surrounding quartz monzonite is parallel to the contact of the intrusive locally. Inclusions of the quartz monzonite in the outer arc of coarse-grained monzonite are rare, but they indicate that some piece-meal stoping has occurred; possibly enough to remove more evidence of parallelism. The author, therefore, favors forceful injection as the most plausible mode of emplacement of the diorite body.

After the diorite body had crystallized, the magmatic pressure of the reservoir which was differentiating at depth was less than that of the surrounding rocks. Fractures, dipping steeply outward, formed along the steeper set of two possible shear fractures, (Billings, 1945, p. 49) and monzonitic magma welled up all around the subsiding block of diorite, forming a monzonite ring-dike encompassing 360°.

Volcanic Vents Near Little Rattlesnake Hill

The volcanic vents (Pl. 1) near Little Rattlesnake Hill, Raymond township, cut quartzite and quartz-biotite schist of the Berwick formation. The structure in the country rock is typical of the region; the beds strike northeast and in most places dip steeply northwest. The igneous bodies transgress the bedding, the attitude of which is unmodified even at the contact. They are chiefly breccia and aphanites, which have been mapped as a single unit and which partly surround cores of gabbro and monzonite.

The coarse-grained rocks are dominantly massive but locally foliation formed by biotite flakes or a lineation produced by hornblende grains was observed.

The igneous mass, which lies 0.5 mile southeast of Little Rattlesnake Hill, has a core of coarse-grained rock, chiefly gabbro, partly surrounded by two arcuate areas of essentially aphanitic rock, composed of keratophyre, quartz latite, and breccia. Exposures of contacts between the igneous body and the country rock are rare.

Three contacts were observed between the coarse-grained gabbro and the aphanitic rocks. The contact between the aphanites and the gabbro strikes N.25°E. and dips 83°SE. in the eastern part of the two knobs and in the southeastern part of the body. This is the only indication that the body has an outward dip. The contact between the gabbro and the breccia in the northeastern part of the body trends N.30°W. and is marked by a depression. The third contact between gabbro and breccia, strikes N.70°E. in the southwestern corner of the body; the dip was not obtainable.

The hill which lies 0.5 mile northeast of Little Rattlesnake Hill is apparently made up entirely of fine-grained, black andesite except for one outcrop of a breccia composed of andesite and rhyolite. Exposures consist of a series of elongate outcrops in which joints are the only observable structural feature. The joints strike parallel to the contours of the hill and commonly dip steeply outward. The relationship to the country rock was not observed.

Little Rattlesnake Hill is underlain by an elliptical body which consists of a core of coarse-grained rock, composed dominantly of hypersthene monzonite, which is partially surrounded on the south by fine-grained rhyolite (possibly containing some keratophyre and quartz latite). A porphyritic border phase of the hypersthene monzonite crops out at the northwest end of Little Rattlesnake Hill between 320 and 340 feet in elevation. The contacts between the hypersthene monzonite and the rhyolite and between the rhyolite and the country rock are well exposed. The strike of the contacts commonly follows the contours of the hill. Although the dip of the contacts ranges from steeply outward to 80° inward, it is dominantly vertical. The igneous body cuts across the bedding of the country rock except at the northwest contact of the rhyolite with the country rock. There the contact follows a bedding plane for a few feet, then cuts across for a few feet and repeats this procedure.

The strike of joints and dikes is generally parallel to the contacts of the body. The dips are variable, although in most places they are steep to vertical.

The igneous bodies in the vicinity of Little Rattlesnake Hill can best be explained as filling volcanic vents. The emplacement of these bodies apparently occurred in two major phases: the intrusive phase and the volcanic phase.

The coarse-grained rocks, gabbro and hypersthene monzonite may have intruded the Berwick formation at different times. The change in rock type from dominantly gabbro (0.5 mile southeast of Little Rattlesnake Hill to andesite 0.5 mile northeast of it, and hypersthene monzonite at Little Rattlesnake Hill suggests successive changes related to the differentiation of magma at depth, or possibly to assimilation of roof rock (granite in part at Little Rattlesnake Hill and quartzite and biotite schist at the other two localities) in separate cupolas. Evidence for the mode of intrusion is meager. Whether the emplacement of the hypersthene monzonite of Little Rattlesnake Hill followed or preceded that of the andesite on Hook Hill is uncertain. The hypersthene monzonite cross-cuts the wall rock, and may have been emplaced by piecemeal stoping.

After time to permit cooling of the hypersthene monzonite to a coarse-grained rock, the volcanic phase started with the eruption at Hook Hill. The vent, formed when the country rock was blown out by the eruption, filled with andesitic magma which chilled to the aphanitic, black andesite now exposed. This was possibly followed by the volcanic eruptions at Little Rattlesnake Hill and at the locality 0.5 miles southeast at Little Rattlesnake Hill. Quartzlatite, ouachitite, and rhyolite were injected along the contacts of the coarse-grained rocks with country rocks. The new accessions of magma disrupted the edges of the previous vents, and at the locality, 0.5 mile southeast of Little Rattlesnake Hill, formed breccias of wall rock and core rock with a fine-grained matrix of the new magma. The new magma crystallized without forming a breccia at Little Rattlesnake Hill.

Joints

Joints and quartz veins are numerous in the rocks of the Mt. Pawtuckaway quadrangle. Observations were made of 183 joints and quartz veins, 92 in the igneous rocks and 91 in the metamorphic rocks. Joints in the rocks of the White Mountain magma series, controlled by local structural conditions, are not considered. The joints occur dominantly in joint sets and locally in joint systems composed of two and rarely three joint sets. The interval between joints is variable, but in most places it ranges from 1 to 10 feet.

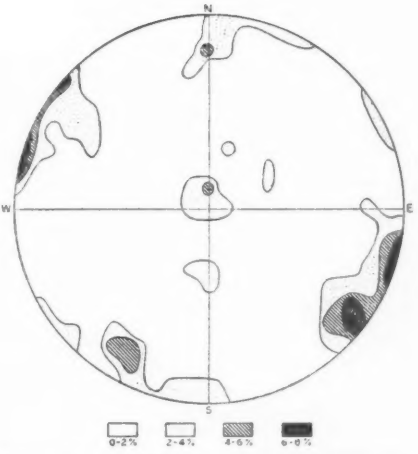

FIGURE 8.—CONTOUR DIAGRAM OF NINETY-TWO JOINTS AND QUARTZ VEINS IN IGNEOUS ROCKS
(Poles of perpendiculars plotted on upper hemisphere of equal area net)

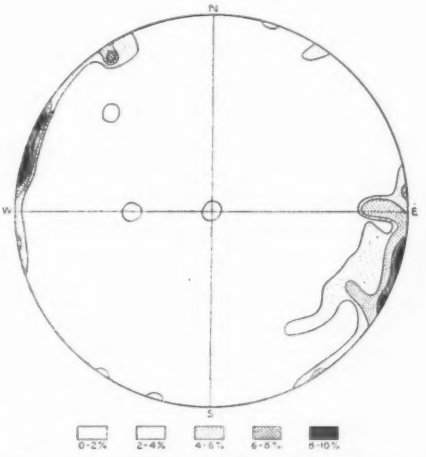

FIGURE 9.—CONTOUR DIAGRAM OF NINETY-ONE JOINTS AND QUARTZ VEINS IN METAMORPHIC ROCKS
(Poles of perpendiculars plotted on upper hemisphere of equal area net)

The contour diagrams (Figs. 8, 9) show concentration in the southeast and northwest which

indicates that most of the joints strike northeast. The maxima near the circumference show that nearly all the joints are vertical or dip steeply northwest or southeast. A concentration in the center of the diagram, which is stronger in Figure 8, is due to horizontal or gently dipping sheeting which is generally parallel to the topographic surface. Deeply eroded massive, plutonic rocks show this feature, caused by tensional forces due to release of load.

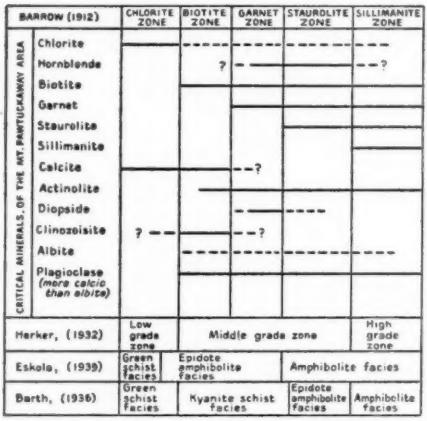

FIGURE 10.—STABILITY OF CRITICAL METAMORPHIC MINERALS AND ZONAL CLASSIFICATION

Solid lines indicate stable minerals; dashed lines indicate metastable or unstable minerals commonly present.

Actinolite refers to that derived from reaction of dolomite and quartz.

Three sets of joints cut the plutonic rocks. The principal set strikes N.20°E. ± 10°; a second set strikes N.70°W. ± 20°; and a third set strikes N.20°W. ± 10°. There is no evidence that these joints are genetically related to the intrusion of the plutonic rocks. They are apparently due to external forces acting on the already consolidated granitic rock. Regional horizontal compression in this area, acting at right angles to the regional trend, was oriented N.70°W. The N.70°W. joints thus appear to be extension joints; the N.20°E. joints are probably release joints; and the N.30°W. joints, shear fractures (Billings, 1942, p. 125).

In the metamorphic rocks the principal set of joints strikes N.20°E. and are vertical. Although they may also be release joints, many of them may be bedding joints, because the bedding throughout much of the area trends N.20°E. and is vertical.

METAMORPHISM

General Statement

The zonal concept proposed by Billings (1937) for indicating intensity of metamorphism in western New Hampshire is equally applicable in southeastern New Hampshire. The metamorphic zones were recognized in the field on the basis of critical minerals present. Changes in zonal boundaries were made as thin-section study revealed unsuspected critical minerals in microscopic quantities. The critical minerals used to determine zones of equal metamorphic intensity are chlorite in the low-grade zone, biotite, garnet, and staurolite in the middle-grade zone, and sillimanite in the high-grade zone (Fig. 10). Actinolite, formed in this area by the metamorphism of rocks containing dolomite and quartz, occurs in the middle- and high-grade zones.

The grade of metamorphism is not a criterion of age, but is due to a variation in intensity of metamorphism in rocks that are progressively younger to the northeast. Mineral isograds mark the boundaries of the northeast-trending intensity zones (Fig. 3). The intensity of metamorphism increases northwest from the southeastern corner of the quadrangle. The chlorite zone coincides in general with the Eliot formation; the biotite zone is superimposed chiefly on the Berwick formation. The garnet zone includes the upper strata of the Berwick formation and parts of the Littleton formation that border the main intrusive body. A narrow belt of micaceous rocks in the Gove member near the base of the Littleton formation contains both staurolite and sillimanite. The sillimanite zone coincides with most of the Littleton formation southeast of the plutonic bodies and in Barrington township. Because of the dominance of quartz-rich rocks, farther southeastward the staurolite isograd is restricted to favorable rocks close to the sillimanite isograd. The intensity of metamorphism increases southeast from the northwestern corner of the quad-

rangle. In this area the intensity zones are disrupted by igneous intrusion and faulting. The Littleton formation along the northwest margin of the quadrangle lies in the biotite zone, which extends to southeast of Saddleback Mountain. On Saddleback Mountain, however, there is an isolated area of more highly metamorphosed rock in which garnet and staurolite occur.

The actinolite isograd (Fig. 3) lies between the biotite and garnet isograds and marks a line west of which actinolitic rocks crop out in the biotite and higher intensity zones.

Cause of Zoning

Barrell (1921), comparing the New England crystalline province with that of the sedimentary rocks in the Appalachians of Pennsylvania, concluded that neither load nor dynamic metamorphism has been as effective in causing metamorphic change as were plutonic intrusives. Billings (1937) attributed the increase in metamorphism toward the southeast in western New Hampshire to the increase in the number and size of the intrusives of the New Hampshire magma series. This principle is well illustrated in the Mt. Pawtuckaway quadrangle. Since the deformation seems to be uniform over the whole quadrangle, changes in intensity of metamorphism are apparently due to aqueous magmatic and pore solutions and changes in temperatures and pressures. Intensity of metamorphism increases northwestward, reaching a peak near the plutonic rocks of the New Hampshire magma series, and apparently decreases northwest from the main mass of the intrusion. The metamorphism is believed to have occurred in one stage, late in the tectonic history of the region.

Equilibrium Relations

The metamorphic rocks of the Mt. Pawtuckaway quadrangle wherever their composition was favorable have essentially attained equilibrium as have rocks in other areas in New Hampshire. However, the rocks of the biotite and higher grade zones contain chlorite. The original chemical composition of the rocks was evidently unfavorable for sillimanite, since it is not abundant.

Parallelism of Isograds and Contacts

The changes rocks undergo as a result of variation in intensity of metamorphism provide the best information about the physical and chemical conditions to which the rocks are subjected. An ideal case would be a series of strata in which metamorphic zones are at an angle to the strike of the beds. Any one bed would appear in more than one metamorphic zone, and the changes in the mineralogy along the bed would indicate what changes had occurred and whether material had been added or removed. In the Mt. Pawtuckaway quadrangle, the isograds in general are parallel to the contacts, which makes it impossible to determine what changes in chemical composition have accompanied the metamorphism.

Occurrence of Minerals

Muscovite or sericite occurs in all the formations and is stable in all the metamorphic zones. It is the dominant micaceous mineral in the Eliot formation and in the Littleton formation around Gulf Hill. It occurs as oriented aggregates of tiny flakes and as large prophyroblasts at an angle to the schistosity. The large porphyroblasts have apparently formed late in, or subsequent to, the deformation. These are most common in the garnet, staurolite, and sillimanite zones.

Muscovite, although stable in all metamorphic zones, took part in the formation of biotite. There is less muscovite in rocks of the biotite zone. However, the apparent increase in the amount of muscovite in the garnet, staurolite, and sillimanite zones may be the result of addition of potash during metamorphism (Billings, 1938). This would account for the late, cross-cutting muscovite.

Plagioclase is abundant only in the schists of the Littleton formation and in the lime-silicate rocks. Albite, characteristic of the chlorite zone in other areas of New Hampshire (Billings, 1937; Kruger, 1946), was not observed in that zone nor was any other plagioclase noted; the original rocks must have had little soda. Albite does occur south of Nottingham Square in Nottingham township and 0.5 mile east of Flint Hill in Raymond township, in the sillimanite and biotite zones respectively. Both occurrences are close to pegmatite injections and hydrothermal veins, and in one specimen the albite showed the curved cleavage of cleavelandite. Albite must have been introduced after the main deformation.

Everyone agrees that albite is the only stable plagioclase in the chlorite zone. There is disagreement, however, as to whether, in the biotite zone and higher, the anorthite content increases with increase in the intensity of metamorphism. Phillips (1930), Turner (1938) and Ambrose (1936) maintain this thesis; Barth (1936) and Billings (1937) believe that above the biotite isograd the anorthite content of a plagioclase depends on the composition of the original sediments. The last-mentioned principle seems applicable to rocks of the Mt. Pawtuckaway quadrangle. The plagioclase in the Mt. Pawtuckaway quadrangle shows no consistent increase in anorthite content with increasing metamorphism. Therefore the anorthite content of the plagioclase of a metamorphic rock is not an indication of the intensity of metamorphism, except that an anorthite content of over 10 per cent indicates the biotite zone or a higher intensity zone.

Chlorite occurs in several ways in the metamorphic rocks in the Mt. Pawtuckaway quadrangle. In the chlorite zone, it is stable and occurs with sericite in flakes and aggregates parallel to the schistosity. In the higher intensity zones, it occurs as relic chlorite, retrograde chlorite, and as chlorite formed at higher intensity conditions than those of the chlorite zone.

Relic chlorite occurs in the center of grains that have been partially metamorphosed to biotite. It is also associated with muscovite, interpreted as a failure to reach equilibrium, since biotite is formed at the expense of chlorite in the presence of muscovite (Harker, 1939, p. 214). Some of the chlorite in the garnet zones may be interpreted in the same way; it is used in the formation of garnet.

Retrograde chlorite can be recognized definitely in only a few specimens. In these the chlorite replaces the fringes of biotite flakes and occurs in veinlets in grains of garnet. A number of the rocks with abundant chlorite are near silicified zones or have themselves been silicified. This chlorite apparently was introduced with the hydrothermal silica that silicified the country rock.

Porphyroblasts of chlorite occur in one outcrop, at an angle to the foliation and with abnormal ultrablue interference colors, indicating that the chlorite is penninite. White (1946) describes similar chlorite which probably formed under the same conditions.

Biotite is abundant except in the chlorite zone. It commonly forms at the expense of chlorite in the presence of muscovite in the biotite zones, and continues into the higher intensity zones. Biotite occurs in scattered patches and aggregates of flakes aligned parallel to the foliation. It is invariably brown, and the grain size is markedly larger close to igneous bodies and even around small quartz veins, lenses and pods. The grain size of the biotite in some of the rocks is 0.05 mm. to 1.0 mm.; around quartz injections the grain size ranges to 5.0 mm.

The optical properties of biotite show no regular change in the successive metamorphic zones (Fig. 5).

The garnets in most of the metamorphic rocks are almandite with some spessartite. In the lime-silicate rocks, the garnet is essentially 70 per cent pyrope + almandite, and 30 per cent grossularite + andradite. Garnet occurs in the garnet zone and zones of higher intensity. Nowhere abundant, more garnets were observed in the rocks on the southeastern slope of Saddleback Mountain than elsewhere. Where in contact with mica, it apparently cuts across the schistosity rather than bowing it out. This indicates that the garnets probably formed after the last movement.

The garnets are porphyroblastic and differ in development of crystal faces, which are dominantly dodecahedrons modified by the trapezohedron. Many of the crystals are fractured and form skeletal growths full of inclusions of other minerals. The better developed crystals are clearer of contamination and fractures.

Staurolite is abundant in the staurolite zone and carries over into the sillimanite zone. Like garnet and much of the biotite, staurolite has a porphyroblastic habit. It occurs scattered and in crystalline aggregates throughout the rocks, but it is most abundant in foliation planes. The largest crystals seen were 20 mm. long, but most of them range from 5 to 10 mm. The prisms lie in the plane of the foliation except where penetration twins form at high angles.

Sillimanite is scarce and was observed in hand specimen in only a few rocks, where it occurred as radiating sheaves and small, elongate bundles of needles. In thin section, smaller amounts of sillimanite occur in scattered needles and bundles.

Lime-silicate minerals were found in the biotite and higher intensity zones. Actinolite, plagioclase, quartz, and biotite constitute the bulk of these rocks. Clinozoisite was observed in one rock in the biotite zone, west of the actinolite isograd, and diopside in another, an inclusion in the quartz monzonite west of Mendums Pond. White (1946) reports an actinolite isograd in the staurolite zone in the Woodsville quadrangle, Vermont, but found that calcareous rocks were not present in the lower metamorphic zones of that quadrangle. In the Mt. Pawtuckaway quadrangle such rocks occur in the lower zones which are thus lithologically favorable for indicating that lower intensity conditions can metamorphose quartz and dolomite to actinolite. The change starts at the actinolite isograd in the

lower part of the biotite zone and continues northwest.

The scarcity of diopside is one of the striking features of the lime-silicate rocks in the Mt. Pawtuckaway quadrangle. Diopside normally forms from actinolite at higher temperature conditions in the presence of solutions. Actinolite forms from dolomite and quartz with excess of $CaCO_3$ and CO_2. If at higher temperature conditions the actinolite reacts with the $CaCO_3$, diopside forms. However, if the $CaCO_3$ is removed from the system by solution, the higher intensity conditions cannot affect the actinolite, which remains stable (Billings, 1937).

Age of the Rocks

General Statement

One of the major problems that prompted study of this area is that of the age of the rocks. A specific determination is impossible with the evidence now at hand, but the Pennsylvanian and pre-Algonkian ages proposed by Katz (1917) for the metamorphic rocks are disproved, and a middle Paleozoic age is inferred.

Fossils were not found in the Mt. Pawtuckaway quadrangle. Age determination must therefore depend upon relationships to dated plutonic rocks and distant fossiliferous areas, and lithologic similarities. Information derived from the study of relationships to dated plutonic rocks is the most reliable and will be discussed first. Correlation with fossiliferous areas is more specific, but is also more open to the possibility of errors. Mineralogical and lithological comparisons in the igneous rocks are reliable, but in the metamorphic rocks, where intensity zones may cross stratigraphic units, that type of evidence is the least reliable of the methods of correlation.

Evidence from the Plutonic Rocks

The rocks of the Pawtuckaway Mountains and the Little Rattlesnake Hill area are similar in mineralogy, lithological characteristics, and structure to rocks of the alkalic White Mountain magma series and are correlated with it. This magma series may be assumed to be post-Devonian because the Moat volcanics belonging to it overlie Devonian metamorphic rocks and rocks of the Devonian New Hampshire magma series with pronounced angular unconformity (Williams and Billings, 1938, p. 1025). The Pennsylvanian Pondville conglomerate, south of the Blue Hills in eastern Massachusetts, contains pebbles of the Blue Hill granite porphyry (LaForge, 1932, p. 35) which is alkalic and is correlated with the White Mountain magma series. Thus the White Mountain magma series is younger than late Devonian and older than early Pennsylvanian. It is classified as Mississippian (?).

In the Mt. Pawtuckaway quadrangle, the rocks of the Mississippian (?) White Mountain magma series are intrusive into the older igneous rocks of the Devonian (?) New Hampshire magma series and into the metamorphic rocks. Rocks of the New Hampshire magma series are also intrusive into the metamorphic rocks.

The metamorphic rocks of the Mt. Pawtuckaway quadrangle are thus not only older than the Mississippian (?) White Mountain magma series, but also older than the New Hampshire magma series, which Shaub (1937), by a lead-uranium ratio determination from an associated pegmatite, indicated is probably late Devonian. The metamorphic rocks are thus Devonian or older.

Correlation with Fossiliferous Areas

Fossils of Pennsylvanian age (Emerson, 1917) were discovered in the Worcester phyllite at Worcester, Massachusetts. Emerson (p. 62–64) quotes David White as follows:—"The specimens at present in hand, though few and very fragmentary, are such as to put beyond question the Carboniferous age of the phyllite at Worcester." A correlation of the Eliot formation with the Worcester phyllite has been assumed in the past as follows: Worcester is approximately 60 miles S.30°W. of the Mt. Pawtuckaway quadrangle, and the regional trend of the country rock in the two areas is aligned; Keith, LaForge, and Katz traced a belt of rock from Worcester, northeastward across New Hampshire and southwestern Maine to Portland, Maine, and attempted a correlation. Beneath the Worcester phyllite is the Oakdale quartzite which can be traced northeastward. In the Lowell area, the Oakdale quartzite is considered the equivalent of the

Merrimack quartzite, and the Worcester phyllite is considered equivalent to the Brimfield schist; farther northeast in New Hampshire, the Kittery quartzite is believed to be the equivalent of the Merrimack quartzite and the Eliot formation the equivalent of the Worcester phyllite. On the basis of this correlation, the Kittery quartzite and the Eliot formation were considered Pennsylvanian (Katz, 1917).

Billings (oral communication) and Currier (1947) believe that there is something wrong with the correlation. There are three possible ways in which the age of the correlated beds could be in error: (1) the fragmentary plant fossils which led David White to make a positive Carboniferous determination may not be reliable; (2) the Worcester phyllite may be Pennsylvanian, but the underlying Oakdale quartzite may be older, with an unconformity between the two formations; (3) there may be an error in tracing the strata northeast.

The fossil evidence for a Carboniferous age of the Worcester phyllite was carefully studied by David White. It is possible, but not probable, that the fossils were incorrectly identified. Although the belt of strata which Katz traced from Worcester, Massachusetts to Portland, Maine is interrupted locally by igneous intrusions, it is probable that the formations can be correlated. The idea that the Oakdale quartzite may be considerably older than the Worcester phyllite furnishes the best possibility of setting up a correlation that would fit evidence from other areas.

Fossils found in the Littleton area in western New Hampshire (Billings and Cleaves, 1934) have been used to determine the age of the Littleton formation as Lower Devonian and the age of the Fitch formation as mid-Silurian (Billings, 1937). As indicated in an earlier part of this report, the Littleton formation has been traced across New Hampshire to the Mt. Pawtuckaway quadrangle. The tracing has been done by detailed mapping; but since several faults are crossed, large intrusions break the continuity, and there may be a facies change from the original sediments, an error is possible, though not probable. The schists in the northwestern half of the quadrangle have therefore been assigned to the Lower Devonian Littleton formation.

The Berwick and Eliot formations are consequently of Lower Devonian age or older; a further clue to their age is obtained from fossils to the northeast. Graptolites classified as of mid-Silurian (Clinton or Lower Wenlock) age were found in Waterville, Maine (Perkins, 1924). Waterville is about 120 miles northeast of the Mt. Pawtuckaway quadrangle. The correlation between the Mt. Pawtuckaway quadrangle and the Waterville slates is as follows: The area of Katz's (1917) work adjoins the Mt. Pawtuckaway quadrangle on the east. Northeast of this area, Fisher (1941), in the Lewiston, Maine area, has made a direct tie to the Waterville slates. The only gap in the continuity is 0° 5′, 43° 45′ to 43° 50′ north latitude, between Katz's and Fisher's work, and this unstudied area is apparently not critical. The correlation with the Waterville slates, more direct that with any other fossiliferous area, is as follows: the mid-Silurian Waterville slates are the equivalent of the Sabattus formation (Fisher, 1941). The Androscoggin and the Pejepscot formations underlie the Sabattus formation. Traced southwesterly, the Pejepscot formation is continuous with the Berwick formation. Fisher assigns the Pejepscot formation to the Cambrian, and the Androscoggin formation to the Ordovician primarily by superposition, but both formations could also be Silurian.

This correlation is clear-cut and probably reliable; it appears therefore that the Berwick formation is Silurian or older. The underlying Kittery quartzite and the overlying Eliot formation are closely related in age to the Berwick and must be Silurian or older. Silurian age is favored because: (1) The Berwick and Eliot formations are directly beneath Lower Devonian rocks; and (2) the formations are lithologically unlike the fossiliferous Cambrian of eastern New England.

At Rowley, Massachusetts (Fig. 1), 25 miles southeast of the Mt. Pawtuckaway quadrangle, "The age of the volcanic complex forming the Newbury Basin, in Essex County, . . . was determined as probably Lower Devonian through the discovery by Keith, in 1915, of marine fossils regarded by Ulrich as probably of that age. . . . " (LaForge, 1932, p. 7). This is little more than a quadrangle away, and it is the

closest fossil locality with which a correlation could be made. However, the rocks in the vicinity of the Newbury volcanic complex are chiefly plutonic, and there is no direct tie to the metamorphic rocks of the Mt. Pawtuckaway quadrangle.

Summary

Thus four lines of evidence are available to determine the age of the metamorphic rocks of the Mt. Pawtuckaway quadrangle: (1) They are intruded by plutonic rocks of probable Mississippian and probable late Devonian age, they are Devonian or older; (2) Correlation with western New Hampshire indicates that the Littleton formation is Lower Devonian, and the Berwick and Eliot formations are of that age or older; (3) Correlation with the Waterville area, Maine, indicates that the Berwick and Eliot formations are middle Silurian or older; (4) Katz, by correlation with the Worcester area, advocated a Pennsylvanian age. The preponderance of evidence, therefore, suggests that the schists assigned to the Littleton formation are Lower Devonian, and the Berwick and Eliot formations are middle Silurian or older.

Summary of Geologic History

Geosynclinal Phase

The geological record in the Mt. Pawtuckaway quadrangle starts with the deposition of calcareous and carbonaceous shale and sandstone of the Silurian (?) Eliot formation. About 6,500 feet of sediments were laid down. Deposition of 7,000 feet of interbedded sandstone and shale and local lenses of dolomitic limestone of the Silurian (?) Berwick formation followed without a break. No break in deposition is evident between rocks of the Silurian (?) and Devonian periods. Shaly sandstones and interbedded shales and sandstones of the Devonian Littleton formation were deposited conformably on the Silurian (?) Berwick formation. Approximately 8,000 feet of sediments were laid down.

A land mass to the east supplied the material to form a total thickness of approximately 20,000 feet of sediments. As most of the sediments are fine-grained, the eastern land mass must have been low, or the source terrane distant from the sea in which they were deposited.

Orogenic Phase

During middle or late Devonian time, the sedimentary strata were involved in tectonic disturbances of the Acadian revolution. The Silurian (?) and Devonian strata were folded to form a homoclinal series on the northwest limb of an anticline. Numerous minor folds from a few inches to a few feet or even larger across are associated with the major fold. A foliation parallel to the bedding formed during the early stages of the orogeny. Lineation, evidenced by crinkles and aligned mica flakes in the foliation, formed contemporaneously with the minor folding. During late stages of the deformation, fracture cleavage developed locally, parallel to the axial planes of the minor folds.

Late in the orogeny, before the last movements had subsided, successive injections of the New Hampshire magma series formed two large bodies of plutonic rocks. The Exeter diorite, the most mafic of the bodies, was probably injected first. The quartz diorite of the Mt. Misery area in the northeastern part of the quadrangle, was intruded in part concordantly and in part discordantly. The quartz monzonite and binary granite cutting across and engulfing the country rock followed closely enough so that their mutual boundaries are gradational. The microcline granite was the last intrusion of the New Hampshire magma series. Locally, it cross-cuts the country rock, but in large, the contact is essentially parallel to the regional trend of the country rock.

Before all of the magma had completely solidified, numerous pegmatite veins, sills, dikes, and lenses were injected.

The metamorphism is partly syntectonic and partly related to the igneous intrusions late in the orogeny. Recrystallization began soon after folding started. The foliation parallel to the bedding formed in that period.

The igneous activity late in the orogeny supplied the high temperatures and solutions which metamorphosed the strata to high-grade

metamorphic rocks. As the intensity of metamorphism decreases away from the igneous body, the rocks farther away from and higher above the magma were subjected to less intense conditions. Most of the rocks crystallized without any significant change in chemical composition.

Mineralizing solutions in great quantities, rest liquors left after the magma had crystallized, permeated the rocks. Penetrating the metamorphic rocks in the Northwood area, they formed migmatites. Veins, pods, and lenses of quartz were emplaced. The quartz probably made room and bowed out the strata by the force of crystallization. The solutions aided recrystallization of the rock minerals to coarser grain size. The metamorphic rocks on Saddleback Mountain were abundantly tourmalinized, and veins of tourmalinized, lime-silicate rock were formed.

Postorogenic History

The igneous rocks of the New Hampshire magma series had crystallized by Mississippian (?) time. Rocks of the White Mountain magma series were then emplaced by cauldron subsidence and ring-dike injection in the Pawtuckaway Mountains, and volcanoes were active in the vents of the Little Rattlesnake Hill area. Numerous dikes formed of rocks of the White Mountain magma series were injected into all the rocks in the quadrangle.

In late Triassic time faults disrupted the rocks, new accessions of silica silicified the metamorphic and igneous rocks along some of the fractures, and diabase dikes were injected.

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SUPPOSED MARGINAL FAULT LINES IN THE SHELF AREA OFF SOME HIGH NORTHERN LANDS

By Olaf Holtedahl

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Abstract

Studies of the shelf topography off Norway indicate that the high northwestern part of the Scandinavian Peninsula is bordered by fractures along which dislocations probably took place during the Tertiary uplift of the land mass. Some of the submarine areas in question are particularly considered, and attention is drawn to the fact that similar relief features are characteristic also of shelf areas off West Greenland and Labrador.

Introductory Statement of Writer's Interpretation of "Marginal Channels"

The writer (Holtedahl, 1933; 1935a; 1940) has drawn attention to some rather striking geomorphological features of the shelf area off the northwest coast of the Scandinavian Peninsula. A number of relatively long and narrow depressions parallel the main trend of the coast line, and, where transverse submarine channels cross the outer part of the shelf, commonly these channels end landward in a relatively steep slope situated in the continuation of the inner slope of the longitudinal depressions (Figs. 2, 3).

It was maintained that these features (like the well-known depression called the Norwegian channel, off the southwesternmost part of Norway) can be explained only by assuming marginal fracture lines along which the topography has been carved out, principally when sea level was lower than at present. It was, furthermore, thought very probable that this supposed tectonic fracturing had a close, causal relation to the fairly recent, mainly Tertiary, uplift of the land mass, which has transformed a more or less peneplained region into a highly dissected land block with heights up to 1000–2000 meters.

In Pleistocene time the ice remodelled the shelf area, changed river channels into trough-like basins, and obscured the old topography through deposition of drift material. This material, largely levelled out and superficially reassorted when the sea invaded the shelf area, now makes up, to an unknown depth, the shallow and flat banks.

Inside the supposed fault zone erosion, according to the writer's view, largely caused by the work of coastal glaciers, has cut into the uplifted land mass.

If these suppositions are valid we should expect similar features in the shelf relief off other glaciated high coastal lands. A study of

charts of the shelf areas of Spitsbergen and especially Greenland revealed that "marginal channels" appear typical of at least parts of these areas (Holtedahl, 1935b; 1936). The writer therefore thinks we are here dealing with features of rather great interest.

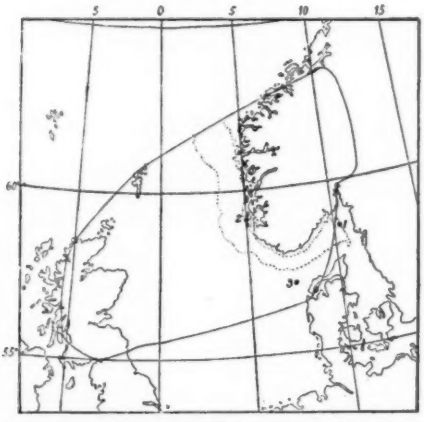

FIGURE 1.—EARTHQUAKE EPICENTERS IN RELATION TO THE NORWEGIAN CHANNEL

(1) Oct. 23, 1904, (2) Jan. 24, 1927, with district of distribution, (3) May 23 and 29, 1929. From Kolderup and Krumbach (1930).

SHEPARD'S VIEW

Recently Shepard in his book *Submarine geology* (1948) has summarized data concerning this broad and many-facetted subject, a subject in which he has made so many important contributions. Shepard refers, however, rather briefly to the existence of longitudinal depressions in shelf areas and evidently does not believe in the theory of fracture lines. He writes (p. 171): "There is little reason to refer to the deeps around glaciated coasts as fault troughs or as drowned river valleys." He thinks the work of glaciers suffices as an explanation. Reference is also made to an earlier publication by Shepard (1931).

Under these circumstances and because the writer's previous publications on the shelf problems may not be readily accessible to American geologists, it seems to be of interest once more to take up the question of marginal fracture lines.

THE NORWEGIAN CHANNEL

This depression, at least 800 km. long and running along a (partly strongly curved) coast made up of old crystalline rocks, is a unique topographical feature and therefore one which must have been formed by extraordinary processes. Among students of the geological history of the Fennoscandian land mass and its borders it is the generally accepted view that this is an example of large-scale crustal fracturing. Even several decades ago, F. Nansen, G. De Geer, and J. J. Sederholm believed that we must assume a zone of faulting. The relatively great and constant width of the depression (60–80 km.) suggests not a single fault line (guiding later erosion) but a subsided area.

One point supporting such a view is the rather abrupt way in which this depression is cut down at its northeastern end (in the Skagerrak area), along a line that probably has close connection with the post-Cretaceous faults in Scania in southern Sweden. The deep depression does not continue northeastward into the Oslofjord district where glacial erosion, owing to the converging movements of the ice masses, must have been especially active.

Of the few more important earthquakes recorded in these regions in this century (Fig. 1) two had their epicenters on the landward slope of the Norwegian channel (Kolderup and Krumbach, 1930). One of the districts of Fennoscandia that has a maximum of seismicity is the coastal area just east of the northern part of this long depression.

Studies on the ash beds in the Eocene "Moler" sediments of Denmark suggest that there has been crustal unrest in the zone under consideration. The Danish geologist S. A. Andersen writes (1937, p. 51–52).

> "Judging from the thickness variations in the tuff layers in Denmark . . . there can hardly be any doubt that the volcanoes lay north of Denmark, and that it was a group or row of volcanoes that stretched along the Norwegian channel. . . ."

OTHER EXAMPLES OF SHELF TOPOGRAPHY OFF THE NORWEGIAN COAST

Figures 2 and 3 show the submarine topography of more northern parts of the shelf area

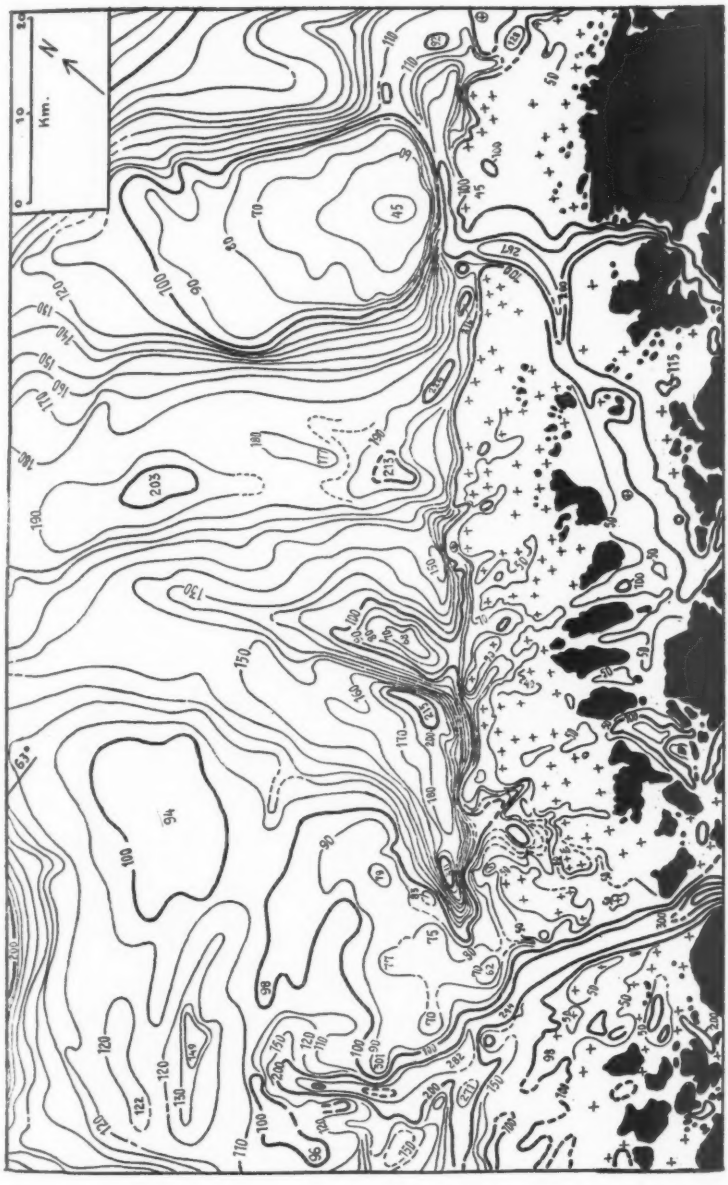

FIGURE 2.—INNER PART OF SHELF AREA OUTSIDE AALESUND-ROMSDAL DISTRICT, NORTHWESTERN SOUTHERN NORWAY
Depth in meters. Crosses mark shoals with depth less than 25 m. Note "Marginal channels" and scarps outside the uneven rocky land mass. (In left part of map area the bank topography suggests former existence of broad curved ice lobes.)

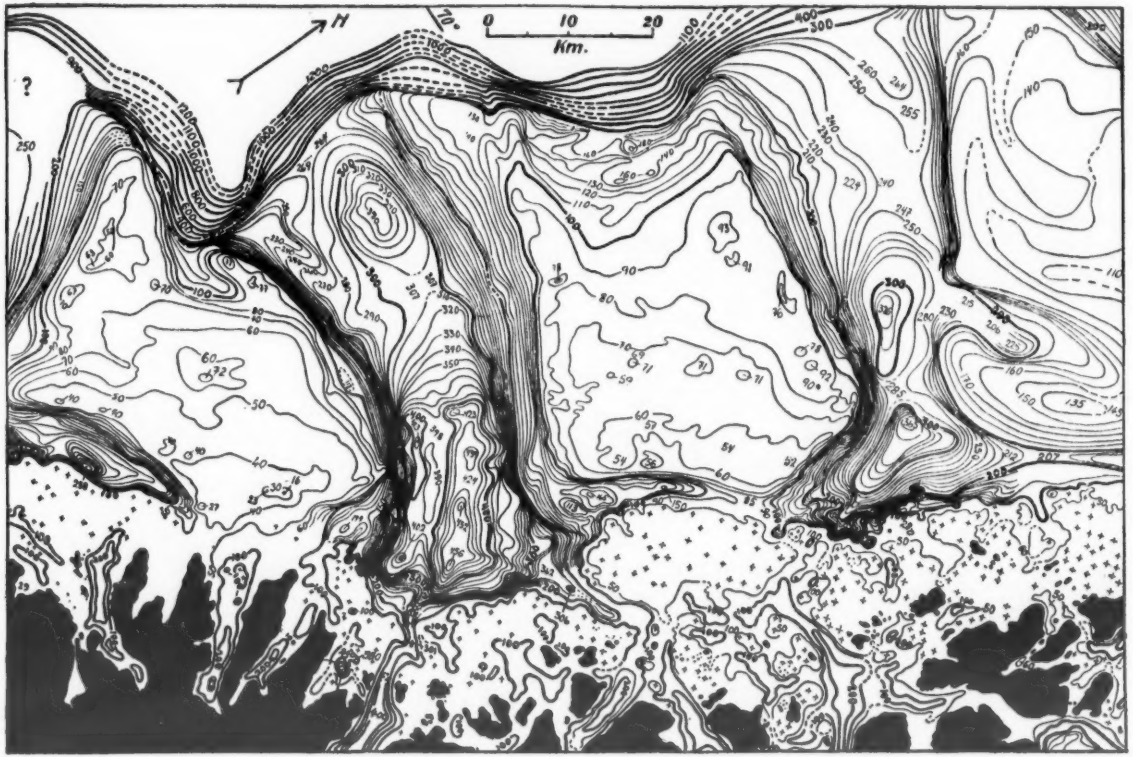

FIGURE 3.—SHELF AREA OUTSIDE TROMSÖ DISTRICT, NORTHERN NORWAY

In lower left corner the island of Senja with heights up to 1000 m. Depths in meters. Same general pattern as in Figure 2 but transverse channels (and bank areas) more marked. Curved trend of supposed fault line in middle part of the picture. From Holtedahl (1935a)

channels (and bank areas) more marked. Curved trend of supposed fault line in middle part of the picture. From Holtedahl (1935a)

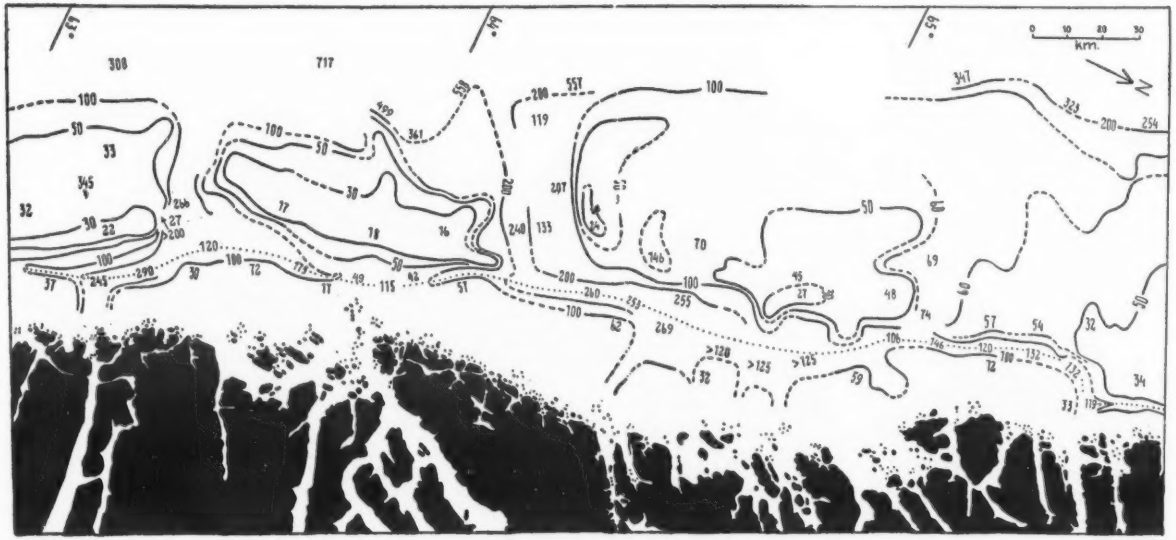

FIGURE 4.—SHELF AREA OUTSIDE A PART OF SOUTHWESTERN GREENLAND

Depths in fathoms. Compare Figure 3. Supposed fracture zone marked by dotted line. Based mainly on (Greenland) chart no. 5, published by Det Kongelige Sökort-Arkiv, Copenhagen 1887.

off Norway—the northwestern part of southern Norway, and a very northern district. It seems to the writer to be a safe conclusion that the development of such striking features as the longitudinal channels and scarps must have been guided by tectonic lines. In both areas no contour lines were drawn. Figure 4 presents contour lines drawn as far as the rather scanty sounding material permits. A comparison of this map with Figure 3 reveals that the two areas show a very striking correspondence in the general pattern of the submarine relief.

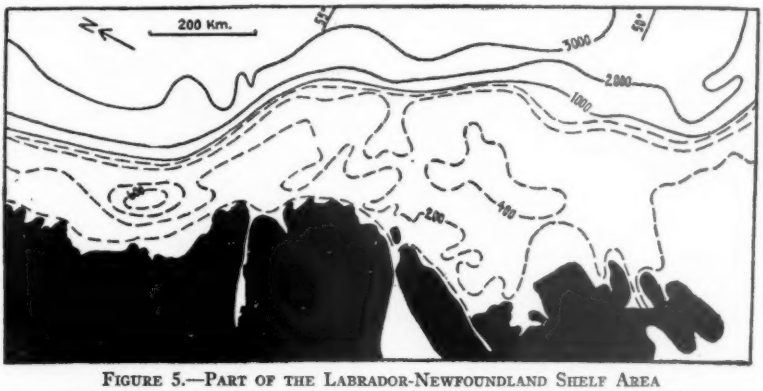

FIGURE 5.—PART OF THE LABRADOR-NEWFOUNDLAND SHELF AREA
Depth in meters. From frontispiece map in Smith, Soule and Mosby (1937).

the rocks of the coastal land are old (pre-Devonian) gneisses and gneiss granites, with varying structural trend.

The transverse depressions are considered the outer part of pre - uplift river valleys, changed into trough basins by subsequent glacial erosion.

Off northernmost Norway the shelf area, which is here part of the wide Barents Sea shelf, is in places very even and flat. Profile 3 of Figure 6 shows how abruptly the submarine part of the rocky land mass rises above the shelf level. Glacial work seems in this area to have been rather ineffective.

In previous discussions the writer has mentioned that the longitudinal zone so well marked in the relief might separate two main types of rock—old hard rocks inside the zone and younger and less resistant ones outside it—but has found such an explanation very improbable. In any case the border line between two such complexes of rocks must have been tectonically determined.

GREENLAND SHELF

In the map accompanying the writer's short note on the West Greenland shelf (1935b)

LABRADOR SHELF

Greenland represents a huge highly uplifted land mass between what Suess termed the Canadian and the Baltic "shields" and separated from both by deep basins which narrow toward the north. A main feature of the two "shields", which are more like shield fragments, is that in the regions bordering the deep sea basins they consist mostly of high land, with old crystalline rocks, the land surface sloping toward the interior of the continents where there is a cover of younger sediments. We have on the American side Baffin Island and Labrador as a pendant to the highly uplifted northwestern part of Fennoscandia (Holtedahl, 1930, maps, Fig. 2).

In a paper by Smith, Soule, and Mosby (1937) the writer has found interesting data concerning the Labrador shelf area and has drawn, from their colored bathymetric map, the contour lines in Figure 5. Of particular interest are profiles of the bottom relief published in the same paper. In Figure 6 some of these profiles are shown, for comparison with some echograms from the North Norway shelf area. These authors have also stressed the

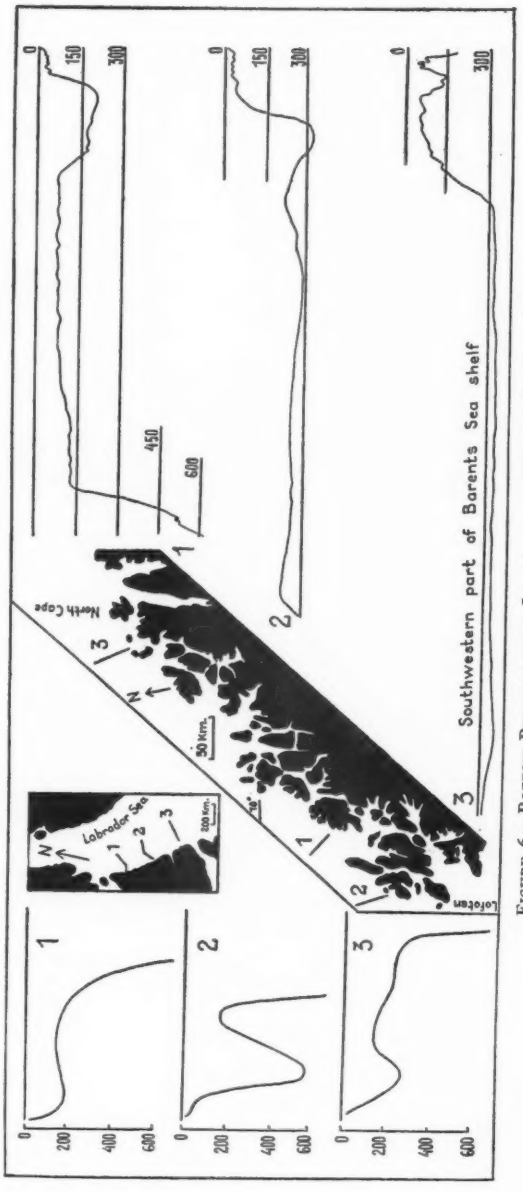

FIGURE 6.—BOTTOM PROFILES FROM LABRADOR AND NORTH NORWAY SHELF
Depth in meters. From Smith, Soule, and Mosby (1937) and Holtedahl (1940).

existence of longitudinal depressions in the Labrador shelf area (p. 3).

A few years ago the geology and geomorphology of the coastal districts of Labrador have been dealt with by an experienced geologist and physiographer who is very familiar with conditions in Fennoscandia—the late Prof. Tanner. He also to some extent discussed the character of the shelf area. In the chapter "The Evidence of the Torngak Mts" he writes (1944, p. 159):

> "The total difference in altitude between the upwarped Palaeic surface and the continental shelf thus amounts to perhaps some 1400–1500 m. max. It is therefore scarcely possible not to assume that the eastern border is formed by a fault zone stretching in the main parallel with the coast."

Concluding Remarks

The main dividing zone between oceans and continents is the continental slope which has come into existence through crust deformations of the first order. The line or zone of supposed marginal fractures dealt with in the present paper is a minor feature, yet one which is important in the discussion of the nature of the border zone between land and sea It is to be hoped that a rapidly growing knowledge of the northern shelf areas soon will throw more light on the problems raised.

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CENOZOIC GRAVELS ON THE RIM OF SYCAMORE CANYON, ARIZONA

BY WILLIAM E. PRICE, JR.

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ABSTRACT

Cenozoic gravels along the rim of Sycamore Canyon are of two types: one containing rounded pebbles derived locally from the Shinarump conglomerate and developed in Pleistocene-Recent time; the other, angular pebbles derived from rocks similar to those exposed today in the Black Hills-Bradshaw Mountains area to the southwest and developed in the Miocene or Pliocene. The angular gravels were laid down, not upon a peneplain as suggested by Robinson, D. W. Johnson, *et al.*, but upon a pediment that extended northeastward from the mountains and covered that area of the Colorado plateau now occupied by Sycamore Canyon.

PURPOSE

Cenozoic gravels of Sycamore Canyon, Arizona, may be of great value in understanding the Cenozoic history of the Grand Canyon region. The possible importance of these deposits led the author, while an associate of the Museum of Northern Arizona, to study them in the summer of 1947. Thanks are given for the facilities provided by the Museum and for the helpfulness of the faculty of the University of Arizona, especially Professor Edwin D. McKee, at whose suggestion the study was undertaken.

LOCATION AND GENERAL GEOLOGY

Sycamore Canyon is in the southwestern escarpment (Mogollon Rim) of the Colorado Plateau, about 25 miles southwest of Flagstaff and 25 miles southeast of Williams, Arizona (Fig. 1). Occupying the canyon, which reaches a depth of 1500 feet, is Sycamore Creek, a southerly flowing tributary of the Verde River. In the upper part of the canyon horizontal or slightly dipping strata of Permian, Triassic, and Cenozoic age are exposed. The Permian formations include, from bottom to top, the Supai, Coconino, Toroweap, and Kaibab formations. Unconformably overlying the Kaibab on the central part of the canyon rim is the Triassic Moenkopi formation. On the eroded surfaces of the Kaibab and Moenkopi formations lie the Cenozoic gravels (Fig. 2). These, in turn, are covered by basalt flows placed by Robinson (1913, p. 30–40) in the First General Period of Eruption (Pliocene). The canyon area has been broken by normal faults, most of which trend northwest-southeast (Fig. 2). The faults are of two groups: one earlier than the basalts, the other later. The pre-basalt faults are responsible for preservation of the soft Moenkopi red beds and the weakly consolidated gravels by having dropped them below the general level of the plateau surface.

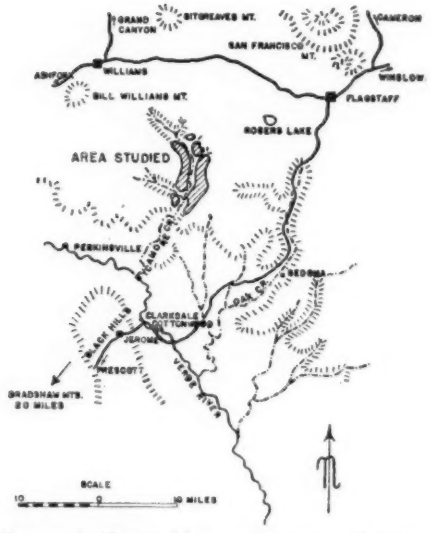

FIGURE 1.—INDEX MAP OF SYCAMORE CANYON

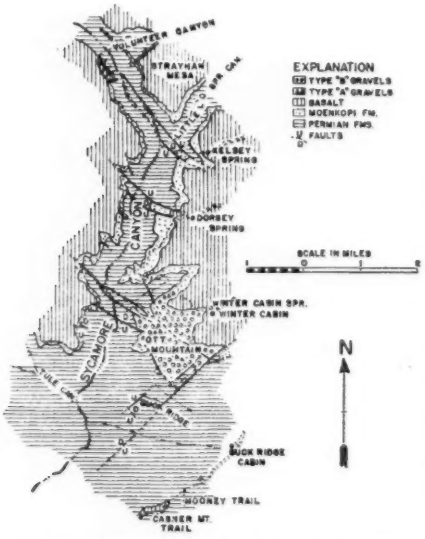

FIGURE 2.—RECONNAISSANCE MAP OF THE RIM ROCKS OF SYCAMORE CANYON

PREVIOUS WORK

Gravels in the walls of Sycamore Canyon were noted by Robinson (1913, p. 30), who considered them Triassic:

> "... the pebbles [are] in the conglomerate at the base of the 'Lithodendron formation' [Chinle formation]. In Sycamore Canyon, many pebbles are distinctly subangular and are composed of gneiss, jasper, and other metamorphic rocks as well as basic igneous rocks of granitic texture. Fairly round cobbles of sandstone and chert up to 8 inches in diameter are less abundant."

DESCRIPTION

Differentiation

Gravels in Sycamore Canyon are of two types: Type A, angular to subangular gravels of varied composition; and Type B, younger, well-rounded gravels mainly of quartzite and chert, derived from the Shinarump conglomerate.

Type A Gravels

Type A gravels occur as patches of disintegrated gravel beds scattered over the surface, and as massive consolidated outcrops.

The gravels in patches lie on slopes below basalt outcrops. Many patches are in heavily forested areas along the canyon rim, so their relationship to other formations is obscure. The patches vary widely in composition and texture. The largest exposure of these gravels is in the Winter Cabin-Ott Mountain area (Fig. 2). Apparently these large exposures are the result of the breakdown of a poorly consolidated conglomerate on the dip slope of a tilted fault block.

Two exposures of Type A gravels were found in place: one in Volunteer Canyon, another along Casner Mountain Trail (Fig. 2). The deposit in Volunteer Canyon, overlying the Moenkopi formation and underlying basalts, is about 7 feet thick and consists mostly of limestone pebbles. The Casner exposure is over 130 feet thick and consists of pebbles of many kinds similar to those of the scattered gravels. This exposure rests on the Kaibab formation and is overlain by basalts.

The scattered gravels are inferred to have been derived from the erosion of large deposits similar to those under the basalt at Volunteer Canyon and along Casner Mountain Trail because: (1) They appear to occupy the same stratigraphic position beneath the basalt; the scattered gravels lie only on surfaces downslope from basalt outcrops and, except in isolated

cases, are not found on outcrops of basalt. The area has been much disturbed by post-basalt faulting, so the exceptions may be explained as having been derived from deposits on upthrown blocks. (2) The degree of roundness and sphericity of the scattered gravels is approximately the same as that of the massive gravels. This suggests a similar degree of transportation under similar conditions. (3) Gravels of identical composition are present in both deposits.

An analysis of these gravels (gravel samples 1–3) is given in Tables 1 and 2.

Recognizable pebble types found in Type A gravels are:

Permian
1. Kaibab formation—several fossils, including *Dictyoclostus bassi*, *Meekella sp.*, and a sponge. Limestone from the alpha member.
2. Coconino sandstone (?).

Mississippian
1. Redwall limestone—pieces of crinoidal limestone and chert nodules.

Cambrian
1. Tapeats sandstone.

Precambrian[1]
1. Mazatzal quartzite.
2. Quartz-tourmaline dike rock—like those near Prescott.
3. "A faint suggestion of some of the metarhyolites north of Jerome.
4. "Diorite porphyry—rather common in Mingus Mountain quadrangle.
5. "Epidotized aplite (?)—not common in Mingus Mountain quadrangle, but similar rock seen near Prescott area.
6. "Quartz porphyry (better, alaskite porphyry)—this facies found south of Jerome.
7. "Diabase (?) or diorite—common south of Jerome.
8. "Quartz porphyry—rock like this occurs as dikes south of Jerome.
9. "Aplitic granite or alaskite—rare in Mingus Mountain quadrangle, similar rocks in small masses found near Humboldt."

Type B Gravels

Type B gravels occur scattered over the surface of outcrops of the Moenkopi formation where the latter forms gentle slopes. Associated petrified wood occurs most commonly as large chips or splinters, but occasionally as small logs, or portions of large logs. The maximum size of the petrified wood observed was about 2½ feet in diameter and 2 feet long. Of the many

[1] Numbers 3–9 kindly identified by Dr. Charles A. Anderson of the U. S. Geological Survey, who is mapping the Mingus Mountain quadrangle, in the Black Hills area.

TABLE 1.—DATA AVERAGES OF GRAVEL SAMPLES
(Samples 1, 2, 3 are Type A; Sample 4 is Type B)

Gravel Sample	Diameter (mm.)			Sphericity	Roundness	Zingg Classification (Per cent)			
	a	b	c			I	II	III	IV
1	47	32	22	0.70	0.44	30	36	20	14
2	44	31	20	.70	.44	36	36	12	16
3	75	52	36	.69	.49	36	32	8	24
4	51	39	28	.76	.62	32	48	4	16

TABLE 2.—COMPOSITION OF GRAVEL SAMPLES
(In per cent)

TYPE A GRAVELS

Composition	Gravel Sample		
	1	2	3
Quartzite	24	28	14
Chert	20	18	16
Sandstone	2		46
Limestone	42		
Andesite	4	8	
Diorite porphyry		10	
Green shale		6	4
Basalt		8	
Granite		8	
Chalcedony	2		4
Felsite		6	
Quartz			6
Epidote-quartz		2	2
Gneiss		2	2
Andesite porphyry			2
Baked shale (?)	2		
Claystone	2		
Diabase			2
Flint		2	
Rhyolite	2		
Rhyolite porphyry		2	
Schist			2

TYPE B GRAVELS

Composition	Gravel Sample 4
Quartzite	74
Chert	24
Chalcedony	2

pieces examined, only one showed evidence of having been transported by water after deposition. Only a single specimen of wood with associated cemented gravel was found, so that few data are available as to the character of the original matrix.

The gravels appear to have been derived directly from the Shinarump conglomerate, and therefore must have come from between the Moenkopi formation and the basalt, although the original gravel beds were not found in place. They may have been washed down from outcrops not now visible, or may have weathered down in place, where the overlying basalt had been removed. Visible contacts between the basalt and the Moenkopi formation are very rare, because the basalt talus tends to cover the easily eroded Moenkopi claystones and siltstones. Exposures of the contact show no trace of Type B gravels, but there can be little doubt of their stratigraphic position.

Type B gravels were seen only in the area around Winter Cabin and Ott Mountain, although they may also be present elsewhere along the rim.

Analyses of the Type B gravels (gravel sample 4) are given in Tables 1 and 2.

Gravel Samples

Samples were taken by roughly outlining a small area, one to several feet in diameter, just large enough to contain approximately 50 pebbles, as judged by eye. From this area exactly 50 pebbles were chosen at random.

Table 1 gives the data averages and Table 2 the composition of the gravel samples collected.

Contact with Underlying Formations

Type A gravels rest both on the Moenkopi formation (in Volunteer Canyon) and on the Kaibab formation (along Casner Mountain Trail).

In Volunteer Canyon no continuous series of exposures is available, but as the deposits range in thickness from a few inches to about 7 feet within short distances the surface upon which they were deposited is irregular, and the gravels occupy depressions in the soft claystones and siltstones of the Moenkopi formation.

The contact between the gravels and Kaibab formation in the Casner Mountain Trail section is covered by talus from the gravels above. Near by, however, is a basalt-capped hill in which no gravels occur between the lava and the limestone, suggesting that the deposit along the Casner Mountain Trail occupies a depression beneath the general level of the top of the Kaibab formation.

Nothing is known about the contact of Type B gravels with the underlying formations.

Origin

The difference between Type A and Type B gravels in composition, roundness, and sphericity indicate different origins. No petrified wood was found in Type A gravels.

The angularity and composition of the Type A gravels demonstrate derivation from an uplifted area, not far distant, composed of bedrock ranging from Precambrian to Permian in age. Such an area is present southwest of Sycamore Canyon in the Bradshaw Mountains and Black Hills. Robinson (1913, p. 30) suggested the possible derivation of the gravels from mountains to the southwest. However, he believed the gravels to be Triassic and concluded that

> "... a portion of southwestern Arizona of unknown extent was, therefore, a land area of sufficient elevation to supply large quantities of waste to the lower-lying country on the north during Triassic time and in all probability during Permian time."

Obviously, then, when these gravels were deposited, the region presented a very different aspect from that of today. The plateau must have been comparatively low, bounded on the west and south by mountain ranges. In the location of the present Verde Valley was a steep pediment or piedmont slope mantled with debris from mountains not far to the south. This slope must have extended northeastward into the Sycamore Canyon area, and Type A gravels are believed to have been deposited on the eroded surface of the Kaibab and Moenkopi formations.

The Type B gravels, on the other hand, resemble in all respects those of the Shinarump conglomerate (Gregory, 1913, p. 424–438). They are believed to have been derived from that formation and may have accumulated through residual weathering or after transportation. Evidence favoring derivation through residual weathering is: (1) Almost no trace of the matrix has been found. (2) Large pieces of petrified

wood are present. (3) The petrified wood, except for one piece, shows no signs of being water worn since petrifaction. (4) The gravels have been found only where they litter the surface in places where the Moenkopi formation is preserved.

Age

The gravels in Sycamore Canyon are probably not of Triassic age as assigned by Robinson (1913, p. 30). The following sequence of events is suggested (Fig. 3).

Step 1.—Uplift and faulting, late Miocene or early Pliocene. The mountain block southwest of Sycamore Canyon, which was the source of the Type A gravels, was elevated with the rest of the plateau but to a greater extent. Block faulting in the Black Hills-Bradshaw Mountains area was noted by Lindgren (1926, p. 12).

Evidence for uplift at this time is from two sources. D. W. Johnson (1909, p. 135–161) found in the walls of Oak Creek Canyon, a few miles east of the area studied here, the profiles of lava-filled canyons, thus indicating that the plateau had been undergoing vigorous erosion before the eruption of the Pliocene basalts. Longwell (1946, p. 832), in his recent studies of the age of the Colorado River, has concluded on the basis of relations between lake deposits and stream cutting tha the plateau was uplifted late in the Miocene or early in the Pliocene.

Faulting doubtlessly accompanied the uplift and, in the Sycamore Canyon area, was responsible for preserving blocks of the Moenkopi formation. Some of these blocks were dropped far enough to preserve also the overlying Shinarump conglomerate. The fact that these faults do not cut the basalt demonstrates that they occurred before the Pliocene eruptions. Similar faulting on the plateau is considered to be late Miocene or Pliocene by Robinson (1913, p. 95), who correlates it with the Basin-Range diastrophism late in the Miocene.

Step 2.—Planation by erosion, and deposition of Type A gravels late in the Miocene or early in the Pliocene (Fig. 3).

Immediately following or accompanying uplift, erosion by streams from mountains to the southwest probably took place in the Sycamore Canyon area, beveling the hard and soft rocks impartially. The surface developed then was probably a pediment, and upon it were deposited sheets of the Type A gravels. Evidence for this are (1) the truncation of strata, (2) the presence of coarse gravels on the truncated beds and beneath the lava, and (3) the presence of other pediment surfaces on the plateau beneath similar old lavas as shown by Babenroth and Strahler (1945, p. 127), Koons (1945, p. 178), and Childs (1948, p. 376–387).

The flat surface beneath the lavas might also be interpreted as (1) a peneplain (Robinson, 1913, p. 91), or as (2) a stripped erosional surface. Robinson's peneplain hypothesis appears to be invalid because: (a) a peneplain would have been developed when the region was at or near sea level, instead of after it had been raised as shown by deep pre-lava canyons in the area; (b) a thick cover of soil would be expected on a peneplained surface, rather than a deposit of coarse gravels. There are also objections to the theory of a stripped erosional surface. On such a surface a steplike topography would be expected, with the harder layers forming cliffs and controlling the base level of erosion while the weaker beds were rapidly removed. In the Sycamore Canyon area all rocks are truncated impartially, and the soft Moenkopi strata are preserved only in downfaulted blocks.

There can be little doubt that the Type A gravels were deposited upon this surface, for they rest on both the Kaibab and the Moenkopi formations and beneath lavas of the First Period of Eruption (Robinson, 1913, p. 38–40). Type B gravels were not developed at this time, however, because: (1) if they had been weathered out of the Shinarump conglomerate then they would be thoroughly intermixed with the Type A gravels; (2) if they had been produced by weathering of hills capped by hard Shinarump, there would have been a stripped erosion surface, rather than a pediment, and field evidence shows that this was not so.

Uplifts in other periods might have been responsible for the development of these gravels. McKee (1937, p. 263) has shown that a land mass in late Triassic time lay southwest of the plateau and was a source of the pebbles in the Shinarump conglomerate. If Robinson (1913, p. 30) is correct in considering the gravels on the rims of Sycamore Canyon to be Shinarump conglomerate, then the uplift responsible for

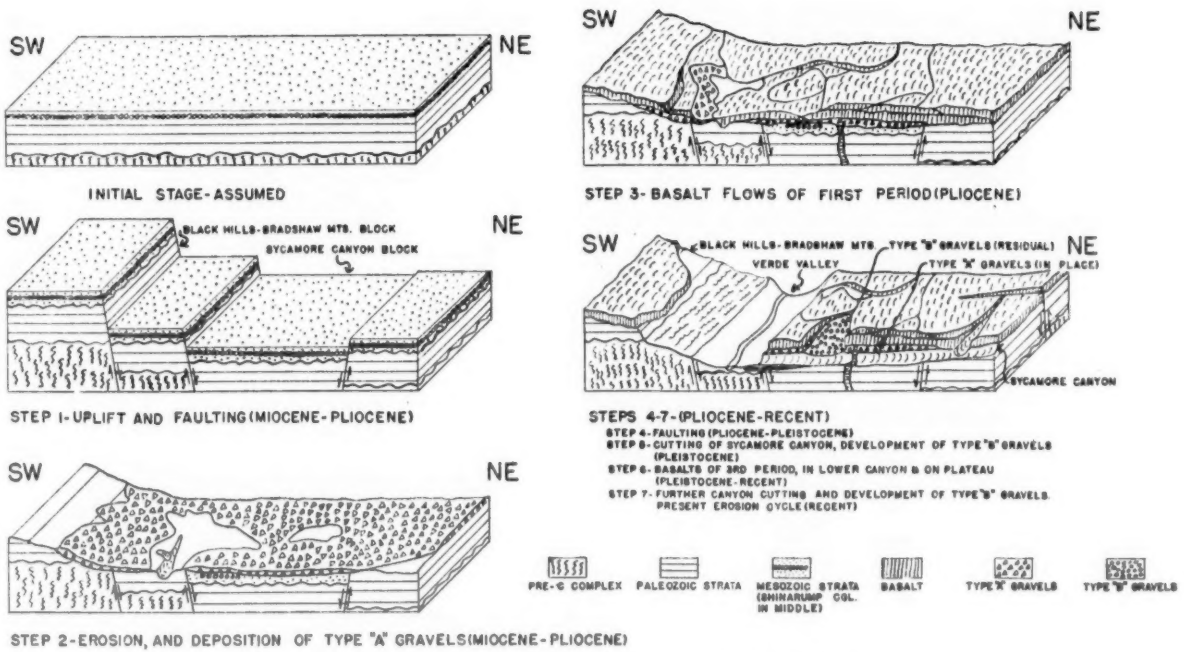
SW
NE
INITIAL STAGE-ASSUMED
BLACK HILLS-BRADSHAW MTS. BLOCK
SYCAMORE CANYON BLOCK
STEP 1-UPLIFT AND FAULTING (MIOCENE-PLIOCENE)
STEP 2-EROSION, AND DEPOSITION OF TYPE "A" GRAVELS (MIOCENE-PLIOCENE)
STEP 3-BASALT FLOWS OF FIRST PERIOD (PLIOCENE)
TYPE "B" GRAVELS (RESIDUAL)
VERDE VALLEY
TYPE "A" GRAVELS (IN PLACE)
SYCAMORE CANYON
STEPS 4-7-(PLIOCENE-RECENT)
STEP 4-FAULTING (PLIOCENE-PLEISTOCENE)
STEP 5-CUTTING OF SYCAMORE CANYON, DEVELOPMENT OF TYPE "B" GRAVELS (PLEISTOCENE)
STEP 6-BASALTS OF 3RD PERIOD, IN LOWER CANYON & ON PLATEAU (PLEISTOCENE-RECENT)
STEP 7-FURTHER CANYON CUTTING AND DEVELOPMENT OF TYPE "B" GRAVELS. PRESENT EROSION CYCLE (RECENT)
PRE-C COMPLEX
PALEOZOIC STRATA
MESOZOIC STRATA (SHINARUMP CGL. IN MIDDLE)
BASALT
TYPE "A" GRAVELS
TYPE "B" GRAVELS

FIGURE 3.—CENOZOIC HISTORY OF THE SYCAMORE CANYON AREA

them must have occurred in the Triassic. But neither gravel type in its present form appears to be Shinarump conglomerate. Type B gravels are not of this formation because they are not consolidated. The Shinarump is recorded as poorly cemented in some places (Darton, 1925, p. 117), but it probably was originally well consolidated in the Sycamore Canyon area because (1) elsewhere in the plateau where it is exposed, it is indurated, and (2) the piece of matrix found at Sycamore Canyon suggests that the Shinarump, if present in the area, is well cemented. The Type A gravels, as their lithology shows, are not the Shinarump conglomerate.

There is no evidence of uplift between the Upper Triassic and the late Miocene or early Pliocene. Orogenic disturbances did produce much clastic material, including conglomerate, in Arizona during the Cretaceous. However, the nearest known Cretaceous deposits are far to the east and northeast, making it improbable that they should also be represented in the Sycamore Canyon area. Because uplift of the plateau area is thought to have occurred in the late Miocene or early Pliocene, they are likely times of gravel development, and there is no reason to resort to a hypothetical earlier uplift.

Step 3.—Basalt flow of First Period in Pliocene time. Evidence that the Sycamore Canyon basalts are of Pliocene age is based upon (1) the fact that they overlie the Type A gravels and therefore are younger, (2) a consideration of later events in the San Franciscan volcanic field, and (3) the time required to erupt the First Period basalts. The Pliocene date of these basalts agrees with that assigned by Robinson. Work by Childs (1948, p. 363) on the geomorphology of the Little Colorado River indicates, however, that these basalts may be very early Pleistocene.

Step 4.—Normal faulting, of Pliocene or Pleistocene age. These faults cut the basalts at Sycamore Canyon and hence are later.

Step 5.—Initial cutting of Sycamore Canyon, probably in the Pleistocene. Evidence that the canyon was cut after the faulting of Step 4 is found in a large tributary on the east rim that follows a post-basalt fault line. During this period Sycamore Canyon was cut down almost to its present depth. The Type B gravels are believed to have accumulated then. Erosion of the protective basalt cover exposed the underlying Shinarump conglomerate in downfaulted blocks, and further erosion, eventually to a point below the base of the conglomerate, produced an eroded Moenkopi surface littered with residual pebbles. McKee (personal communication) states that such effects are produced today in many parts of the Painted Desert by weathering of the Shinarump conglomerate. This hypothesis explains why the Type B gravels occur in concentrated patches and are not intermixed with Type A gravels.

Step 6.—Basalt flows in Recent time. These basalts, found in small patches on the lower walls near the mouth of Sycamore Canyon, have been cut through by Sycamore Creek, but erosion has not proceeded very far below the base of the flows. These basalts can probably be correlated with some of the recent basalts in the San Franciscan volcanic field.

Step 7.—Continued cutting of Sycamore Canyon in Recent time. Probably the continuance of erosion produced more Type B gravels.

In summary, the evidence indicates that the Type A gravels are of Miocene-Pliocene age and the Type B gravels are residual deposits of Pleistocene-Recent age.

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